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THE PHYSICAL ENVIRONMENT OF THE MACKENZIE VALLEY, NORTHWEST TERRITORIES: A BASE LINE FOR THE ASSESSMENT OF ENVIRONMENTAL CHANGE

Edited by
L.D. Dyke and G.R. Brooks

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Rotational landslide in glacial Lake Mackenzie deltaic sediments, near mouth of Mountain River, Northwest Territories. Photograph by J.M. Alysworth, GSC 2000-038

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SUMMARY

If the CO_2 content of the atmosphere continues to rise and the predicted response of climate to this increase is correct, then the Mackenzie valley will experience one of the highest rises in mean annual air temperature for any region in Canada. This possibility calls into question the response of permafrost and the behaviour of ice-rich sediments which typify this region. Although ice-rich permafrost inherently resists climate warming, the potential for thawing to disrupt communities and industrial and economic activity warrants an evaluation of terrain sensitivity so that the consequences of climate warming in the Mackenzie valley can be anticipated. This report is divided into five sections that provide a basis for understanding how climate change will physically affect the Mackenzie valley.

The physical nature and climate of the Mackenzie valley is presented in the first section, with a focus on the distribution, type, and ice content of surficial materials in the region and the geological processes that produced them. The second section deals with reconstructions of past climate. These reconstructions provide indications of climate variability that can be compared with predicted changes. In the third section, permafrost throughout the Mackenzie valley is characterized, emphasizing the relationship between ground temperatures and surface environment. A documentation of ground ice along a north-south transect through part of the Mackenzie valley is included to supplement the ground-ice data presented in the first section. Presentations of active-layer thickness and the distribution of permafrost along shorelines complete this section. These two sections cover aspects of permafrost distribution that are strongly dependent on factors that may be as influential as climate change. In the fourth section, the two most active processes currently changing the landscape are described: river flow and landslides. The possible impact of climate change on these processes is also discussed. The report is completed with the fifth section on predicted changes in permafrost thickness and distribution under specified amounts of warming. The following paragraphs summarize the content of the report.

The present climate of the Mackenzie valley is described in terms of the factors affecting permafrost and landscape-forming processes. Physiography has a considerable influence on regional climate. The Cordillera to the west shields the Mackenzie valley from an influx of warm maritime Pacific air during the winter, thus favouring the residence of cold arctic air masses. This almost uniform winter cold is occasionally broken by the invasion of low-pressure cells from the Gulf of Alaska. During summer, the movement of warm air masses northward from the south makes the region warm for its latitude. A change in frequency of winter low-pressure cells and summer warm air masses could significantly alter both seasonal and annual

SOMMAIRE

Si la teneur en CO_2 de l'atmosphère continue de croître et qu'elle modifie comme prévu les conditions climatiques, la vallée du Mackenzie connaîtra l'une des hausses les plus marquées de la température annuelle moyenne de l'air de tout le Canada. Cette hypothèse soulève la question du comportement du pergélisol et des sédiments riches en glace qui caractérisent cette région. Bien que le pergélisol à haute teneur en glace présente une résistance inhérente au réchauffement climatique, il est possible qu'un dégel perturbe les lieux habités et les activités industrielles et économiques. Pour cette raison, il est justifié d'évaluer la sensibilité des terrains afin d'être en mesure de prévoir les conséquences du réchauffement climatique dans la vallée du Mackenzie. Le présent rapport est divisé en cinq sections contenant des renseignements de base pour comprendre comment le changement climatique modifiera physiquement la vallée du Mackenzie.

La première section donne un aperçu de la physiographie de la vallée du Mackenzie et du climat qui y règne, en mettant l'accent sur la répartition, le type et la teneur en glace des matériaux superficiels, ainsi que sur les processus géologiques qui les ont produits. La deuxième section traite des paléoclimats. Des reconstitutions permettent de comparer les variations du climat avec les changements prévus. Dans la troisième section, le pergélisol de toute la vallée du Mackenzie est décrit, en mettant l'accent sur le lien qui existe entre les températures du sol et l'environnement de surface. Des données sur la glace de sol le long d'un transect nord-sud, traversant une partie de la vallée du Mackenzie, y complètent les données sur la glace de sol présentées dans la première section. De plus, on y indique les épaisseurs du mollisol et la répartition du pergélisol le long des berges de cours d'eau. Ces deux sections couvrent les aspects de la répartition du pergélisol qui dépendent dans une large mesure de facteurs qui peuvent être aussi importants que le changement climatique. Dans la quatrième section, les deux processus actuels qui transforment le plus le paysage sont décrits : l'écoulement fluvial et les glissements de terrain. Les répercussions possibles du changement climatique sur ces processus sont également traitées. La cinquième et dernière section porte sur les changements prévus de l'épaisseur et de la répartition du pergélisol dans le cas d'un réchauffement important. Les paragraphes qui suivent résument le contenu du rapport.

Le climat actuel de la vallée du Mackenzie est décrit en fonction des facteurs qui perturbent le pergélisol et des processus géomorphologiques. La physiographie influe considérablement sur le climat régional. La Cordillère qui s'élève à l'ouest de la vallée du Mackenzie empêche l'air maritime chaud du Pacifique d'affluer en hiver, d'où une stagnation des masses d'air froid en provenance de l'Arctique. Ce froid hivernal quasi uniforme est parfois déstabilisé par l'arrivée de cellules de basse pression en provenance du golfe de l'Alaska. Durant l'été, le déplacement du sud au nord de masses d'air chaud contribue à rendre le climat de cette région relativement chaud, compte tenu de la latitude. Toute modification de la fréquence des cellules de basse pression en hiver et des masses d'air chaud en été pourrait avoir des répercussions considérables sur les températures saisonnières et annuelles à l'échelle régionale. Les

regional temperatures. Thus, mean annual temperatures can also be changed when the frequencies of synoptic events which disrupt these patterns are changed.

Climate is the fundamental environmental factor that determines if ground temperature will remain at or below 0°C throughout the summer. If it does, then permafrost, by definition, is present. This condition is presently favoured throughout the Mackenzie valley, but the existence of permafrost locally is controlled by vegetation, drainage, snow cover, and soil type. This is true for areas in which mean annual air temperatures are warmer than about -6°C, because the environmental factors listed above tend to dampen the penetration of winter cold. The thermal regime of permafrost throughout the Mackenzie valley is also controlled by the same factors that determine permafrost presence. For example, in the southernmost area, permafrost is associated closely with peatland character. In such cases, permafrost up to a few metres thick generally underlies peat bogs that may be elevated slightly and thus dry out in summer. This condition maximizes the insulating properties of the living vegetation mat during the warmest part of the year. Fens, characterized by a higher water table, are unfrozen. In the Fort Simpson area (mean annual air temperature: -3.7°C), permafrost reaches a thickness of 20 m and underlies roughly 30% of the terrain. Its presence in this area is also dependent on forest type. For example, spruce forests are most effective at intercepting snowfall and reducing the depth of snow on the ground, thus inducing colder ground temperatures and deeper permafrost. In the Norman Wells area (mean annual air temperature: -6.0°C), permafrost reaches a thickness of 60 m and underlies about 80% of the terrain. At Inuvik (mean annual air temperature: -9.5°C), permafrost is virtually ubiquitous and thicknesses in excess of 100 m are common; however, in the Mackenzie Delta, permafrost is generally less than 100 m thick because of warmer ground temperatures associated with dense, snow-trapping, willow growth and the warming effect of both migrating distributary channels and abundant lakes.

The glacial history of the Mackenzie valley is fundamental to understanding the distribution of surface sediments. Much of the physical character of the Mackenzie valley is derived from a series of temporary glacial lakes that were impounded by retreating ice during the waning of Laurentide glaciation. Sand, silt, and clay sediments up to 60 m thick were deposited in these lakes. These sediments are partially or completely frozen, and together with adjacent glacial deposits, commonly contain ice. This ice, as well as the particularly thick ice masses in the Tuktoyaktuk Peninsula, constitutes one of the most interesting, but troublesome, aspects of the region. Thawing greatly reduces the strength of these frozen sediments and results in a volume decrease, particularly where excess ice in the form of lenses and layers is present. This may induce

températures annuelles moyennes pourraient donc varier si la fréquence des événements synoptiques qui perturbent ces schémas était modifiée.

Le facteur environnemental fondamental qui maintient le sol à une température égale ou inférieure à 0 °C durant tout l'été est le climat. Si cette température se maintient, il y aura, par définition, présence de pergélisol. Les conditions favorables à la présence de pergélisol existent à l'heure actuelle dans toute la vallée du Mackenzie, mais la végétation, le drainage, le manteau nival et la nature des sols peuvent jouer un rôle déterminant par endroits. Ainsi, dans les régions où les températures annuelles moyennes de l'air sont supérieures à -6 °C environ, les facteurs environnementaux énumérés ci-dessus pourront empêcher la formation de pergélisol en contrecarrant la pénétration du froid hivernal. Les facteurs qui régissent le régime thermique du pergélisol dans toute la vallée du Mackenzie sont les mêmes que ceux qui déterminent la présence du pergélisol. C'est ainsi que dans l'extrême sud de cette région, le pergélisol est étroitement associé aux tourbières. Un pergélisol d'une épaisseur pouvant atteindre quelques mètres s'étend généralement dans les tourbières oligotrophes (bogs), lesquelles peuvent se soulever légèrement et, par conséquent, s'assécher en été. Cette situation accroît de façon maximale les propriétés isolantes du couvert végétal vivant durant la période la plus chaude de l'année. Les tourbières minérotrophes (fens), caractérisées par une nappe phréatique plus élevée, ne gèlent pas. Dans la région de Fort Simpson (température annuelle moyenne de l'air : -3,7 °C), le pergélisol atteint une épaisseur de 20 m et s'étend à environ 30 % du terrain. Sa présence dans la région dépend également du type de forêt. Les forêts à épinette sont, par exemple, les plus efficaces à intercepter les chutes de neige et à réduire l'épaisseur du manteau nival. Ainsi, les températures du sol sont plus basses et le pergélisol peut se prolonger à plus grande profondeur. Dans la région de Norman Wells (température annuelle moyenne de l'air : -6,0 °C), le pergélisol atteint une épaisseur de 60 m et occupe environ 80 % du territoire. À Inuvik (température annuelle moyenne de l'air : -9,5 °C), le pergélisol est pratiquement omniprésent et dépasse souvent 100 m d'épaisseur. Cependant, dans le delta du Mackenzie, le pergélisol présente généralement une épaisseur de moins de 100 m. Cette plus faible épaisseur du pergélisol est attribuable à des températures du sol plus élevées, lesquelles résultent de la présence de saules denses retenant la neige, ainsi que de l'effet de réchauffement associé à des défluents à parcours variable et à un nombre élevé de lacs.

Afin de comprendre la répartition des sédiments superficiels, il faut connaître l'histoire glaciaire de la vallée du Mackenzie. La physiographie de cette vallée a été façonnée par une série de lacs glaciaires temporaires dont les eaux étaient retenus par des glaciers en recul durant les stades finaux de la glaciation laurentidienne. Jusqu'à 60 m de sable, de silt et d'argile se sont déposés dans ces lacs. Ces sédiments sont partiellement ou entièrement gelés et, tout comme les sédiments glaciaires voisins, ils contiennent habituellement de la glace. C'est cette glace, ainsi que les amas de glace particulièrement épais dans la péninsule de Tuktoyaktuk, qui constituent l'un des aspects les plus intéressants, mais les plus inquiétants de cette région. Le dégel diminue beaucoup la résistance des sédiments gelés en en réduisant le volume, en particulier s'ils contiennent de la glace excédentaire sous forme de lentilles et de couches. Le dégel peut déclencher des glissements de terrain, augmenter l'alluvionnement des cours d'eau et réduire ou éliminer la

landslides, increase sediment contribution to rivers, and reduce or eliminate the ability of earth materials to support buildings and transportation facilities. Although ice contents are highly variable, they are generally related to sediment textures. Silt and clay tend to have the highest ice contents because of the high capillarity which promotes ice-lens growth; amounts in excess of 50% are possible. Therefore, glacial-lake sediments (silt and clay) and fine-grained tills (sediments deposited by glaciers) are most likely to be ice rich, whereas fluvial sediments (sand and gravel) are least likely to have an ice content that exceeds the pore volume.

Proxy climate indicators, namely pollen in lake-bottom sediments, tree rings, prehistoric sand-dune activity, and ground-temperature distributions in boreholes can be used to reconstruct climate prior to the instrument record. Different amounts of detail are available, depending on the precision associated with a particular indicator. Between 10 000 and 6000 years BP, summer temperatures were probably 1 to 3°C warmer than at present and conditions were drier. Dune fields developed after the retreat of the Laurentide ice sheet, when barren areas of glacial sediments were exposed to wind erosion. These dune fields may have been stabilized by a cooler, wetter period after 6000 years BP. More detailed climatic reconstruction is available for the last 1000 years based on studies of tree-ring chronologies. Tree growth is one of the major pieces of evidence for the occurrence of the Little Ice Age in the Mackenzie valley, during which the mean annual air temperature was probably a maximum of 2 to 3°C cooler than at present. Ground-temperature measurements in boreholes also show evidence of warming over the last 50 to 200 years. This is evident in areas where shallow ground temperatures remain out of equilibrium with the temperature at the ground surface.

Modelling that takes into account the insulating effect of vegetation and snow cover is used to predict the response of permafrost to continued climate warming under different warming scenarios. For a 40 km^2 area surrounding Fort Simpson, permafrost is predicted to virtually disappear under a 2°C increase in mean annual air temperature (Wright, 1995). For the same increase in temperature at Norman Wells, it is predicted that permafrost will be reduced to about half its present extent. It should be noted that these are equilibrium predictions, in which permafrost is fully adjusted to a mean annual temperature that is different from that of the present. Therefore these models give no information about how quickly these changes will take place or on the severity of impacts which may be associated with melting ground ice.

To predict the rate of thaw, modelling has also been carried out for specific locations where ground-ice contents are known (Geo-Engineering (M.S.T.) Ltd., 1995). A 4 to 5°C increase in mean annual air temperature due to a doubling of atmospheric CO_2 is predicted

capacité des matériaux meubles à supporter des bâtiments et des installations de transport. La teneur en glace, très variable, est généralement liée à la texture des sédiments. On observe les teneurs les plus élevées dans les dépôts de silt et d'argile étant donné que la plus forte capillarité qu'offrent ces matériaux favorise la croissance des lentilles de glace; la proportion de glace peut dépasser 50 %. C'est ainsi que les sédiments glaciolacustres (silt et argile) et les tills à grain fin (sédiments déposés par les glaciers) sont le plus souvent riches en glace, tandis que les sédiments fluviatiles (sable et gravier) sont moins susceptibles de contenir un volume de glace qui dépasse le volume des vides interstitiels.

Les indicateurs climatiques de substitution, comme le pollen récolté dans les sédiments lacustres, les cernes de croissance des arbres, l'évolution préhistorique des dunes de sable et les diagraphies de la température du sol peuvent servir à reconstituer le paléoclimat. Selon la précision de l'indicateur utilisé, on peut obtenir des données plus ou moins détaillées. Dans la période de 10 000 à 6 000 ans qui nous précède, les températures estivales étaient probablement plus chaudes de 1 à 3 °C qu'elles ne le sont aujourd'hui, et les conditions plus sèches. Des champs de dunes se sont formés après le recul de l'Inlandsis laurentidien, au moment où des sédiments glaciaires dépourvus de couvert végétal sur de vastes étendues ont été exposés à l'érosion éolienne. Une période plus froide et plus humide, après 6 000 ans, a peut-être stabilisé ces champs de dunes. La dendrochronologie (datations établies à partir des cernes de croissance des arbres) permet de reconstituer de façon plus détaillée le climat qui régnait au cours du dernier millénaire. Les cernes de croissance sont l'un des principaux indicateurs du Petit âge glaciaire dans la vallée du Mackenzie. Durant cette période, la température annuelle moyenne de l'air était probablement plus basse qu'aujourd'hui, de 2 à 3 °C au maximum. Les mesures de la température du sol dans des sondages indiquent également un réchauffement au cours des 50 à 200 dernières années. Ce réchauffement est évident dans les zones où les températures du sol à faible profondeur ne sont pas en équilibre avec la température enregistrée à la surface du sol.

Les modèles qui tiennent compte de l'effet isolant de la végétation et du manteau nival servent à prévoir la réaction du pergélisol à un réchauffement climatique continu par l'application de différents scénarios. Dans un secteur de 40 km^2 entourant Fort Simpson, on prévoit que le pergélisol disparaîtrait pratiquement si la température annuelle moyenne de l'air devait augmenter de 2 °C (Wright, 1995). Si on applique les mêmes prévisions à Norman Wells, l'étendue du pergélisol diminuerait de moitié. À noter que ce sont des prévisions basées sur des réactions d'équilibre, où le pergélisol réagirait dans son ensemble à une température annuelle moyenne différente de celle d'aujourd'hui. Ces modèles ne renseignent donc aucunement sur la rapidité de ces changements ou sur l'importance des répercussions qui pourraient être associées au dégel de la glace de sol.

Afin de prévoir le taux de dégel, des modèles ont été produits pour des endroits bien définis où la teneur de la glace de sol est connue (Geo-Engineering (M.S.T.) Ltd., 1995). Le modèle de circulation générale du Centre canadien de la modélisation et de l'analyse climatique prévoit que, dans la vallée du Mackenzie, la

for the Mackenzie valley by the general circulation model of the Canada Centre for Climate Modelling and Analysis. If it is assumed that this doubling takes place in 50 years and that the climate responds on this time scale, then an air temperature trend can be constructed. This warming over 50 years would eradicate permafrost with average ice contents in the Fort Simpson area and cause substantial thinning from the top down of permafrost in the Norman Wells area, but would only increase the thickness of the active layer (the surface layer which thaws and refreezes each year) in the Tuktoyaktuk area. Changes in snow cover are taken into account in the model, as a decrease and increase relative to current amounts, because of the uncertainty of the manner in which snow accumulation will respond to an increase in temperature. If snow cover did decrease, then ground temperatures would remain cool because of decreased insulation in winter.

Ground-surface conditions controlling the transfer of heat between the air and the ground are probably the most difficult factors to account for in modelling permafrost formation or disappearance. The predictions given by the models are indications of what may happen if climate change consists of only an increase in mean annual air temperature. The actual surface conditions at a site, including snowfall, may be different and more complex than those included in the model. Although the tendency of permafrost to thaw, and the relative responses of permafrost at different general locations are likely correct, the predicted depth of thaw given for a particular site is only an estimate, and probably is inaccurate, especially where real conditions differ from those incorporated in the model.

Processes related to thawing permafrost and ground ice will also respond to climate warming. Most disruptive will be landslides and loss of ground-bearing capacity. Because the response of these processes depends on so many factors, accurately predicting the timing for slope failure or the amount a given area will subside is difficult. Instead, sensitivity to thawing has been determined so that the tendency of a given process to respond to climate warming can be estimated. The stability of river banks in frozen sand and silt deposits will be markedly reduced by a 5 to 10 m reduction in the thickness of frozen sediments; however, if it takes 50 years for this thinning to take place, then distinguishing a failure by this mechanism from ongoing fluvial toe erosion will be difficult. Attributing landslides to climate change may be easier with forest-fire-related active-layer flows. If climate change results in more burning, then more fire-related slides will occur.

The present warming trend can reasonably be expected to continue into the twenty-first century, although the accuracy of the predicted warming is arguable. Considering that permafrost in the southern part of the Mackenzie valley is scattered over wide areas and is at most only a few metres thick, the consequences of this warming will not be severe. Physical

température annuelle moyenne de l'air augmentera de 4 à 5 °C si la quantité de CO_2 atmosphérique devait doubler. Si l'augmentation s'étalait sur 50 ans et que le climat était modifié en conséquence durant cette période, la tendance se dégageant de la variation de la température de l'air pourrait être établie. Un tel réchauffement étalé sur une période de 50 ans éliminerait le pergélisol à teneur en glace moyenne de la région de Fort Simpson et il réduirait considérablement l'épaisseur du pergélisol dans la région de Norman Wells, du haut vers le bas de la coupe gelée. Dans la région de Tuktoyaktuk, par ailleurs, le seul effet prévisible serait une augmentation de l'épaisseur du mollisol (la couche de surface qui dégèle et regèle à chaque année). Dans ce modèle, on tient compte des variations de l'épaisseur du manteau nival selon qu'elle diminue ou augmente par rapport à celle d'aujourd'hui, puisque nous ne savons pas quel sera l'effet d'une augmentation de température sur les accumulations de neige. Si l'épaisseur du manteau nival diminuait, les températures du sol s'abaisseraient par suite de l'amincissement de la couche d'isolation en hiver.

Les conditions de la surface du sol qui régissent le transfert de chaleur entre l'air et le sol sont probablement les facteurs les plus difficiles à prendre en compte dans la modélisation de la formation ou de la disparition du pergélisol. Les modèles donnent des indications sur ce qui peut se produire si le changement climatique n'est causé que par une augmentation de la température annuelle moyenne de l'air. Les conditions qui existent réellement à la surface, incluant les précipitations nivales, peuvent différer et être plus complexes que celles incluses dans les modèles. Même si les prévisions quant à la tendance au dégel du pergélisol et à la réponse relative du pergélisol à différents endroits généraux sont probablement exactes, il ressort que l'évaluation de la profondeur prévue du dégel à un endroit donné ne constitue qu'une estimation et est donc probablement inexacte, surtout si les conditions réelles diffèrent de celles incorporées dans le modèle.

Les processus liés au dégel du pergélisol et à la présence de glace de sol réagiront également au réchauffement climatique. Les processus les plus perturbateurs sont les glissements de terrain et la perte de capacité portante du sol. Étant donné le grand nombre de facteurs qui entrent en jeu, il est difficile de prévoir avec précision le moment où se produira la rupture d'un versant ou l'importance du tassement du sol en un endroit donné. La détermination de la sensibilité d'un sol au dégel vise à faciliter l'estimation de la réponse d'un processus donné au réchauffement climatique. La stabilité de berges de rivières gelées qui se composent de sable et de silt diminuerait de façon marquée si l'épaisseur des sédiments gelés accusait une réduction de 5 à 10 m. Cependant, si l'amincissement du pergélisol s'étalait sur 50 ans, il serait difficile de distinguer une rupture due à ce mécanisme de l'érosion fluviale normale par sapement. On aurait plus tendance à attribuer les glissements de terrain à un changement climatique si ceux-ci étaient liés à un fluage du mollisol attribuable à des feux de forêt. Si le changement climatique augmentait le nombre des feux de forêt, le nombre de glissements dus à ces feux serait plus élevé.

La tendance au réchauffement que l'on observe aujourd'hui devrait se maintenir au cours du prochain siècle, même si l'on peut contester l'exactitude des prévisions. Étant donné que le pergélisol dans la partie sud de la vallée du Mackenzie est réparti de façon irrégulière et qu'il mesure tout au plus quelques mètres d'épaisseur, les conséquences du réchauffement ne seront pas graves. Les processus géomorphologiques, qui interfèrent avec les activités

effects on the terrain, which interfere with land use, are unlikely to occur until thicker, ice-rich permafrost regions along the middle and upper parts of the discontinuous permafrost zone (north of Fort Simpson) are warmed sufficiently to promote widespread thaw. In the meantime, short-term events, such as particularly warm summers and forest fires, must be included in any assessment of environmental change that is likely to affect permafrost. Other geomorphic processes, such as fluvial erosion and deposition, which are intrinsic to the landscape, will continue regardless of foreseeable climate changes. Presently these processes are causing local modifications to the distribution and temperature of permafrost which can be as pronounced as any changes associated with predicted climate warming.

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Geo-Engineering (M.S.T.) Ltd.
1995: Potential impact of global warming in permafrost in the Mackenzie Valley; Geological Survey of Canada, Open File 3017, 224 p.

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d'aménagement des terres, ne devraient pas se produire tant que les régions où le pergélisol est plus épais et riche en glace, le long des parties intermédiaire et supérieure de la zone de pergélisol discontinue (au nord de Fort Simpson), ne subiront pas un réchauffement suffisant pour favoriser le dégel. Entre-temps, des phénomènes de courte durée, comme des étés particulièrement chauds et des incendies de forêt, doivent être pris en compte dans toute évaluation d'un changement environnemental susceptible de modifier le pergélisol. D'autres processus géomorphologiques, comme l'érosion fluviale et l'alluvionnement, intrinsèques à ce type de milieu, se poursuivront quels que soient les changements climatiques prévus. À l'heure actuelle, ces processus modifient ici et là la répartition du pergélisol et sa température; ils peuvent être aussi prononcés que les changements associés au réchauffement climatique prévu.

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Introduction

L.D. Dyke[1] and G.R. Brooks[1]

Dyke, L.D. and Brooks, G.R., 2000: Introduction; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks· Geological Survey of Canada, Bulletin 547, p. 7–10.*

Abstract : There is an increasing acceptance in the scientific community that the present trend of global warming is at least partially driven by rising concentrations of greenhouse gases (particularly CO_2) in the atmosphere because of anthropogenic activities over the past few centuries. Modelling of the future effects of increasing levels of greenhouse gases indicates that warming will be greatest in polar areas and that the Mackenzie valley, Northwest Territories, may experience the highest rise in mean annual temperature for any region in Canada. Because of the widespread presence of permafrost and accompanying ice-rich Quaternary sediments, the Mackenzie valley is particularly sensitive to climate warming. This Bulletin is the report of the Mackenzie Integrated Research and Monitoring Area project which was initiated to assess the sensitivity of landscape-altering processes in the Mackenzie valley to climate warming and to produce regional maps of permafrost and ground-ice distribution.

Résumé : Les membres de la communauté scientifique s'accordent de plus en plus pour dire que le réchauffement planétaire est, du moins en partie, causé par la libération croissante de gaz à effet de serre (en particulier de CO_2) dans l'atmosphère par les activités anthropiques et ce, depuis plusieurs siècles. La modélisation des effets résultant d'une hausse des concentrations des gaz à effet de serre indique que le réchauffement sera plus marqué dans les régions polaires et que la vallée du Mackenzie (Territoires du Nord-Ouest) sera la région du pays qui pourrait connaître le plus fort accroissement de la température annuelle moyenne. Étant donné la présence répandue de pergélisol dans la vallée du Mackenzie, et celle concomitante de sédiments quaternaires riches en glace, cette région est particulièrement sensible au réchauffement climatique. Le présent bulletin est en fait le rapport du projet de recherche et de surveillance intégrées du Mackenzie, lequel visait à évaluer les effets du réchauffement climatique sur les processus géomorphologiques actifs dans la vallée de ce fleuve et à dresser des cartes régionales de la répartition du pergélisol et de la glace de sol.

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DESCRIPTION AND PURPOSE

The Mackenzie valley is the lowland paralleling the Mackenzie River along its course from Great Slave Lake northwest to the Mackenzie Delta on the Beaufort Sea (Fig. 1). The distinctiveness of the Mackenzie valley as a physiographic feature varies, but is most pronounced where the river flows between the Mackenzie and Franklin mountains. Elsewhere it is a broad, undulating valley which rises gradually away from the Mackenzie River. Economically, the Mackenzie valley is a vital water, land, and air transportation corridor for communities along the Mackenzie River and Beaufort Sea coast, as well as for the hinterland on both sides of the river. Large hydrocarbon deposits that represent important national energy reserves are present at Norman Wells and in the Mackenzie Delta–Beaufort Sea area. Although the natural-resource potential of the region is incompletely known, it, combined with developing tourism and an increasingly independent aboriginal population, will maintain the economic importance of the region for the foreseeable future. It is imperative that the attributes of the region that make it particularly sensitive to development be understood and considered, as demands continue for construction and maintenance of transportation, industrial, and domestic facilities.

Permafrost is an important characteristic of the Mackenzie valley. It is patchy in the south, no more than a few metres thick, and closely associated with peat accumulations in wetlands. Along the Beaufort Sea coast, permafrost is virtually continuous and in many areas more than 500 m thick. Because of the widespread presence of permafrost and accompanying ice-rich Quaternary sediments, the Mackenzie valley is a region that is sensitive to climate change. Permafrost in the southern Mackenzie valley is especially susceptible to climate warming because of the large area in which it is thin and at temperatures close to 0°C. Predicted increases in air temperature will result in thawing and land subsidence.

In response to concerns about the effects of global climate change, the Mackenzie valley is one of three regions selected for the Geological Survey of Canada's 'Integrated Research and Monitoring Area' (IRMA) program. The objective of this program is to identify regions of Canada with geological

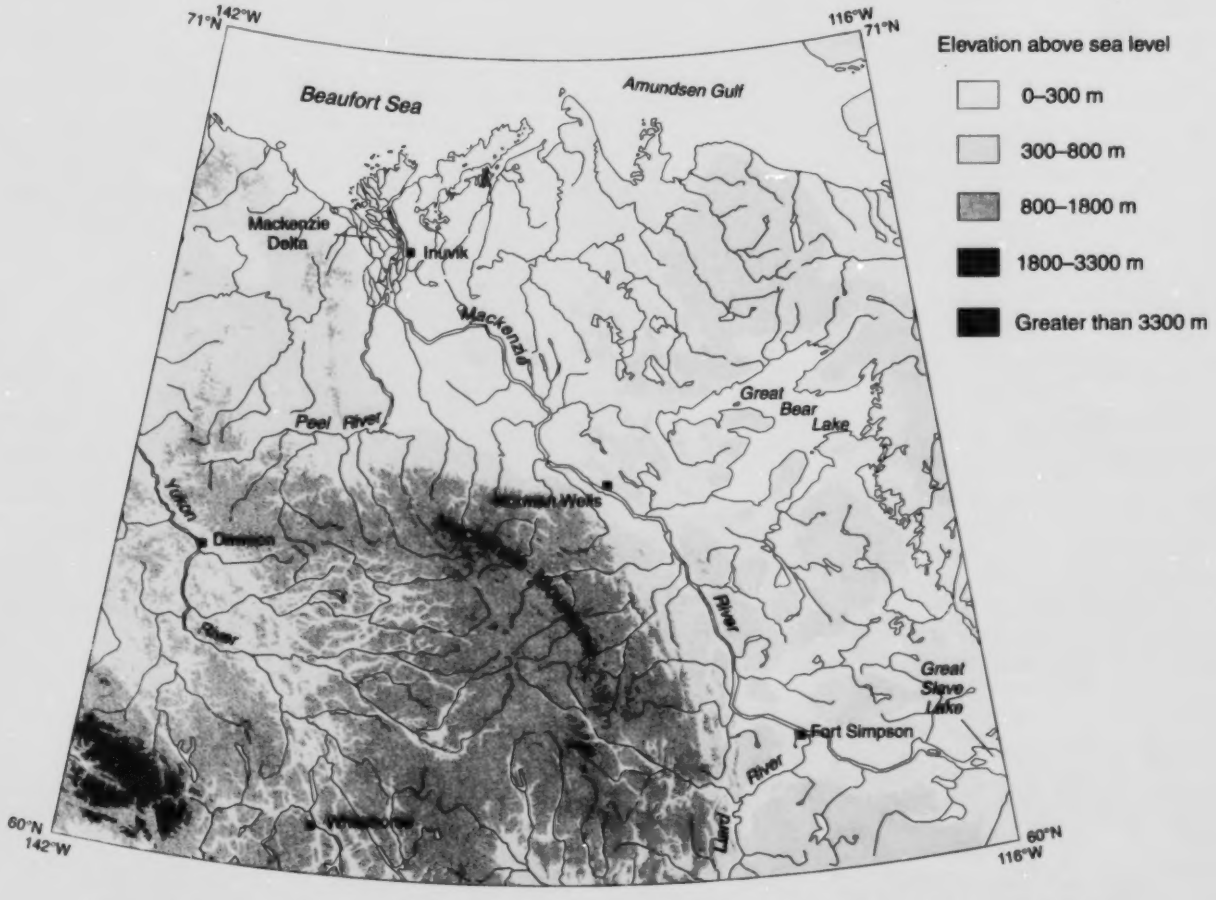

Figure 1. *Physiographic map of the Mackenzie valley area. Topographic features and rivers are labelled in more detail in Figure 1 of Dyke (2000).*

conditions or processes especially sensitive to climate change, and to determine the response of these conditions and processes to this change. Results from the Mackenzie IRMA contributed to the *Mackenzie Basin Impact Study* (Cohen, 1997), a comprehensive investigation by Environment Canada into the implications of climate change for a wide variety of social, economic, biological, as well as geological aspects of the entire Mackenzie River drainage basin. Southern Saskatchewan–southern Alberta (Palliser Triangle) and the Fosheim Peninsula in northwestern Ellsemere Island (High Arctic) were the other areas selected for the IRMA program. Specifically, the purpose of the Mackenzie IRMA was to reconstruct the climate since the end of the Laurentide Ice Age (about 13 ka BP) based on the relationship between climate and vegetation, to produce maps of permafrost and ground-ice distribution, and to determine the sensitivity of landscape-altering processes to climate warming.

The Geological Survey of Canada has been collecting data on permafrost, the character of frozen ground, and geomorphic processes in the Mackenzie valley since the early 1970s. These studies were initiated in response to engineering and environmental concerns arising from hydrocarbon exploration, pipeline proposals, and highway construction. Later, activities focused on monitoring of the Inter-Provincial pipeline built between Norman Wells, Northwest Territories and Zama, Alberta. The Mackenzie Valley IRMA represents a convenient umbrella under which this geoscience information can be assembled to provide a comprehensive summary of permafrost character, past climate change, and response of processes controlled by ground-ice thaw to climate warming. The Mackenzie valley IRMA report, then, is partly a synthesis of information about permafrost in the Mackenzie valley. The report is written for a more general rather than specialized scientific audience, with the purpose of presenting each topic in a manner useful for general knowledge, educational purposes, early stages of project planning, and as background for continued scientific research. The report also serves as a reference for use in engineering design and land-use planning in areas where information about geology and permafrost is required.

CLIMATE WARMING

During the last few decades, anticipation has been growing internationally that increasing concentrations of greenhouse gases in the atmosphere will change the climate in ways detrimental to social and economic patterns. Considering that this bulletin is a response to this concern, it is appropriate to compare predictions of climate change with known climate variation over similar or longer intervals of time in the past.

Abundant data demonstrate that global climate has warmed during the past century. Temperature records extending back to 1860 reveal that the global mean surface air temperature has increased by 0.5 to 0.7°C since that time (Houghton and Woodwell, 1989). The temporal pattern of increasing temperature is not constant, but rather consists of warming and cooling cycles at intervals of several decades. Nonetheless, the long-term trend is one of net global warming with the ten warmest years since 1880 all having occurred in the 1980s and 1990s (Hansen et al., 1996). Corresponding with this warming, alpine glaciers are generally retreating, sea levels are rising, and climatic zones are shifting (Houghton and Woodwell, 1989). All these changes exemplify the environmental impact of global warming.

Although recent global climatic warming is unmistakable, the mechanisms driving the change are poorly understood. Over the past millennium, global temperatures were warmer than present in the interval from roughly AD 1000 to 1300 (Medieval Warm Epoch) and cooler from roughly AD 1450 to 1890 (Little Ice Age). It can be argued that the present warming trend represents the natural transition from a cool to a warm period, a process that has occurred several times in the past; however, the present warming period also corresponds to a significant rise in some atmospheric 'greenhouse' gases (CO_2, CH_4, N_2O, and CFCs and other halocarbons) through human activities, primarily the burning of fossil fuels since the start of the industrial revolution, and from deforestation, animal husbandry, landfills, and rice agriculture (Houghton et al., 1996). Greenhouse gases have an important effect on the Earth's energy budget and contribute to the warming of the atmosphere by trapping infrared radiation emitted from the Earth's surface that would otherwise be transmitted to space. This absorbed radiation is later re-emitted, some of which is transmitted towards Earth and contributes to the warming of its surface. The entire process is commonly called the 'greenhouse effect'. Major increases in any one of the greenhouse gases result in a greater trapping and re-emittance of infrared radiation by the atmosphere, thus leading to higher surface temperatures.

Of the greenhouse gases influenced by the activities of man, CO_2 currently contributes more to the greenhouse effect than all of the other gases combined (excluding water vapour) (Houghton et al., 1996). Ice-core data from Antarctica reveals a close correlation between temperature and CO_2 levels extending over the past 160 000 years (Houghton and Woodwell, 1989); however, this is a coarse correspondence, with no indication of which of the two parameters causes the other to change. Furthermore, the correspondence does not appear to exist at smaller time scales, such as the last few centuries, during which the cool interval known as the Little Ice Age occurred. The Little Ice Age represents the coolest interval during the last 10 000 years. Although warming since the Little Ice Age is attributable to an increase in CO_2 which began about the same time, the onset of the Little Ice Age fails to correspond with any decrease in CO_2. This discrepancy weakens the argument that all subsequent warming is due to the increase in CO_2 concentration which started with the industrial revolution; however, it should be noted that an extensive analysis of climate data resulted in the Intergovernmental Panel on Climate Change (IPCC) reporting, "Our ability to quantify the human influence on global climate is currently limited because the expected signal is still emerging from the noise of natural variability... Nevertheless, the balance of evidence suggests that there is a discernable human influence on global climate." (p. 5, Houghton et al., 1996)

Another climate-controlling factor that corresponds more closely to climate reconstructions is the amount of incoming energy from the sun. This energy or solar radiation can be inferred from ^{14}C values in tree rings because ^{14}C is produced by the interaction of ^{14}N and radiation in the upper atmosphere. Changes in the amount of solar radiation seem to correlate better with our records of past climate (Lean et al., 1995). There appears to be a return to what has been the 'normal' climate throughout much of the last 10 000 years. Recent increases in CO_2 have probably contributed to this warming, but a more thorough understanding of factors that initiated events like the Little Ice Age is required before the role of CO_2 can be fully understood.

Over the past several decades, considerable effort has been devoted to modelling the future effects of increasing levels of CO_2. Models suggest that the concentration of atmospheric CO_2 will increase 35 to 170% by 2100, depending on economic, demographic, and policy factors, as well as assumptions about processes that transform and remove gases from the atmosphere (Houghton et al., 1996). Predictions from climate models indicate that warming will be greatest in the polar regions. Estimates from different models vary, but all agree that a rise of 3–5°C will occur in the Mackenzie valley if the atmospheric CO_2 content doubles by 2050 (Smith and Cohen, 1993). Although predictions of future warming produced by general circulation models may be considered extreme, a continuation of the net warming trend established since the end of the nineteenth century is reasonable to expect.

Recent climate change in the Mackenzie valley is evident from instrumental records that, for a few locations, extend back to the 1890s. These data indicate that mean annual air temperatures have increased 1.7°C over the last century, the highest increase for any climatic region in Canada (Environment Canada, 1995); however, this change may not be representative of the entire Mackenzie valley because of the scarcity of data for such a large area. If it is representative, this warming amounts to a 100–200 km northward shift of isotherms that represent mean annual air temperature. The southern limit of permafrost in the northern prairie provinces has also moved north; however, this response has lagged behind increases in air temperature because of the insulating effect of the vegetation cover (Halsey et al., 1995). Also, recent climate warming is not ubiquitous in the Canadian north, for example, a decrease in average annual air temperature over the same time interval has occurred in the eastern Arctic and permafrost is thickening in northern Quebec due to this local cooling.

ACKNOWLEDGMENTS

The idea of assessing the implications of climate change for the Mackenzie valley was originally developed by P.A. Egginton and was pursued under the federal government's Green Plan initiative. The editors are in debted to B.R. Pelletier for continuous encouragement, advice and technical review throughout the assembly of the volume.

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Glacial history of the Mackenzie region

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Duk-Rodkin, A. and Lemmen, D.S., 2000: Glacial history of the Mackenzie region; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 11–20.*

Abstract: The present landscape of the Mackenzie valley is primarily a result of the last continental glaciation which covered most of the area about thirty thousand years ago (30 ka BP). This glaciation resulted in rearrangement of the preglacial drainage system, including redirection of much of the Mackenzie River drainage to the Arctic Ocean. Meltwater channels established along the edge of the retreating ice sheet became parts of present Mackenzie River tributaries. Lakes were impounded by the ice as it retreated rapidly into the present Mackenzie valley area about 13 ka BP. These glacial lakes received silt and clay accumulating to thicknesses of several tens of metres. Today these sediments include some of the more problematic ice-rich materials with which engineers and construction personnel must deal. By 10 ka BP, the Mackenzie valley was ice-free. Northward tilting of the land continuing to empty and shrink the remaining glacial lakes to their present size.

Résumé : Le paysage actuel de la vallée du Mackenzie a été principalement façonné au cours de la dernière glaciation continentale, au moment où les glaces ont recouvert presque toute la région il y a quelque 30 000 ans (30 ka). Cette glaciation a réorganisé le réseau hydrographique préglaciaire en redirigeant notamment une grande partie du réseau du fleuve Mackenzie vers l'océan Arctique. Les chenaux d'eau de fonte qui se sont formés en bordure de l'inlandsis en retrait sont devenus des affluents du fleuve Mackenzie actuel. Des lacs ont été créés vers 13 ka par la retenue des eaux de fonte dans l'actuelle vallée du Mackenzie, au front des glaces en retrait rapide. Dans le fond de ces lacs glaciaires, du silt et de l'argile se sont accumulés sur plusieurs dizaines de mètres d'épaisseur. Ce sont ces sédiments qui constituent les matériaux riches en glace qui causent le plus de soucis aux ingénieurs et aux constructeurs. Avant 10 ka, la vallée du Mackenzie était totalement dépourvue de glaces. L'inclinaison vers le nord de la surface du sol a permis la vidange des lacs glaciaires et leur rétrécissement jusqu'à leur taille actuelle.

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INTRODUCTION

The landscape of the Mackenzie region is largely the result of glacial activity. Since the Late Pliocene, northwestern Canada has been affected by several montane and continental glaciations. (The term 'montane' glaciation/glaciers is used here to refer to valley glaciers that occupied the Mackenzie Mountains, and is distinct from the 'Cordilleran Ice Sheet' which refers to the continuous carapace of ice that occupied the Selwyn Mountains on the western side of the continental divide.) Montane glaciers extended to the foothills as piedmont lobes, except during the last glaciation when only valley glaciers were present. The earliest montane glaciation of the region, recorded in the Mackenzie Mountains (Fig. 1), is considered to be of Late Pliocene age (Duk-Rodkin et al., 1996). Deposits of the montane glaciations were truncated by the Late Wisconsinan Laurentide Ice Sheet. There is no clear record of how many times continental (Laurentide) ice

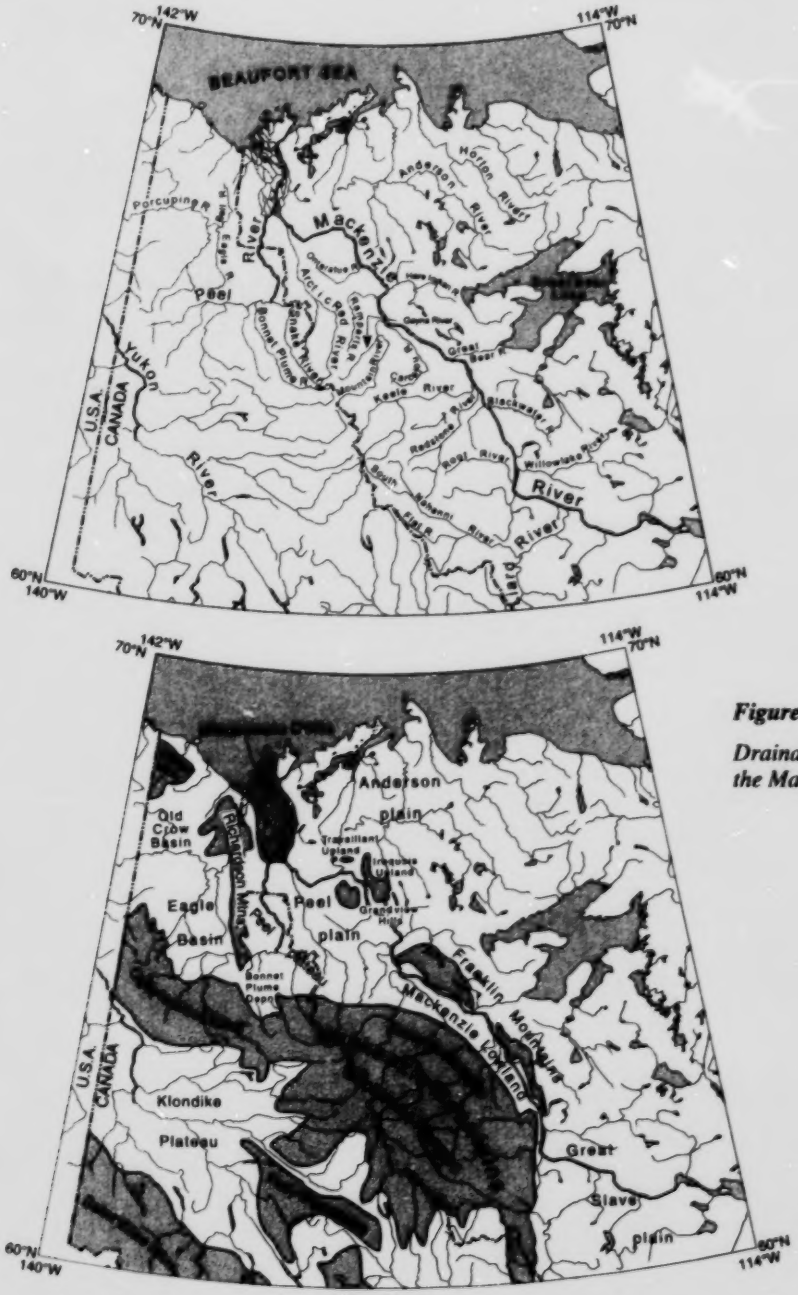

Figure 1.

Drainage and major physiographic features of the Mackenzie region.

advanced over the Canadian northwest, although at least five glaciations are recorded on Banks Island (Barendregt et al., 1998). The stratigraphic and geomorphic records of the Richardson and Mackenzie mountains indicate that the Late Wisconsinan Laurentide Ice Sheet was the only continental glacier to reach these areas. Prior to glaciation, the Mackenzie River valley had a very different landscape.

During the Late Wisconsinan, continental and montane glacier advances were not synchronous. The Laurentide Ice Sheet reached its maximum about 30 ka BP (thousands of radiocarbon years before present) while montane ice was building near the continental divide in the Backbone Ranges of the Mackenzie Mountains (Fig. 1). The Laurentide Ice Sheet was already retreating when montane glaciers reached their maximum extent; although at a few localities the two coalesced. The glaciers in the mountains retreated rapidly, and by 9 ka BP occupied approximately the same areas as modern glaciers while the margin of the Laurentide Ice Sheet still lay near the eastern boundary of the northern Interior Plains.

PREGLACIAL DRAINAGE

In preglacial time, drainage extended from the mountains eastward across the entire northern Interior Plains. The modern Mackenzie River formed much later, primarily as a result of glacial erosion and drainage diversion. The region presently drained by the Mackenzie River formerly contributed to two major drainage systems: one entering the north Atlantic (the Bell River system; Fig. 2), the other entering the Arctic Ocean (Porcupine River and Anderson River; Duk-Rodkin and Hughes (1994)). The Porcupine River drained across the Richardson Mountains to the Mackenzie Delta area starting in the Early Eocene (Duk-Rodkin and Hughes, 1994). River terraces in the mountains (Duk-Rodkin and Hughes, 1992) are correlative with alluvial deposits on the eastern side of the delta. The Peel and Snake rivers were tributaries to the Anderson River as indicated by topographic reconstruction of the region, as well as by clasts found in the Anderson River Tertiary deltaic sediments.

Advance of the Late Wisconsinan Laurentide Ice Sheet diverted these preglacial drainage systems. The Porcupine River was directed to the Pacific Ocean via the Yukon River. Drainage as far south as the Athabasca and Peace rivers was diverted northwestward into the Arctic Ocean. A series of ice-marginal and proglacial channels associated with retreat of the Late Wisconsinan Laurentide Ice Sheet integrated drainages to form the Mackenzie River. Most of the rivers occupying the foothills and plains of the Mackenzie region (e.g. Peel, Snake, Ramparts, Mountain, Carcajou, and others) follow former ice-marginal channels for at least part of their modern courses. Hence the Mackenzie River, Canada's largest drainage system, is entirely postglacial in age and the present interglacial is unique in terms of the amount of fresh water delivered to the Arctic Ocean from North America.

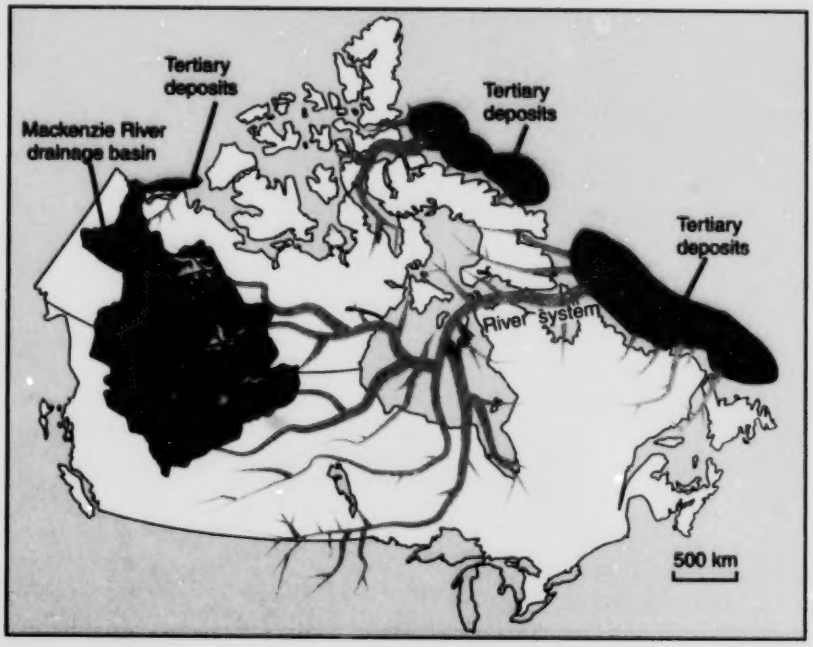

Figure 2. *Late Pliocene to Pleistocene drainage systems (Duk-Rodkin and Hughes, 1994). The modern Mackenzie River flows perpendicular to this drainage, and is a product of glacial diversion.*

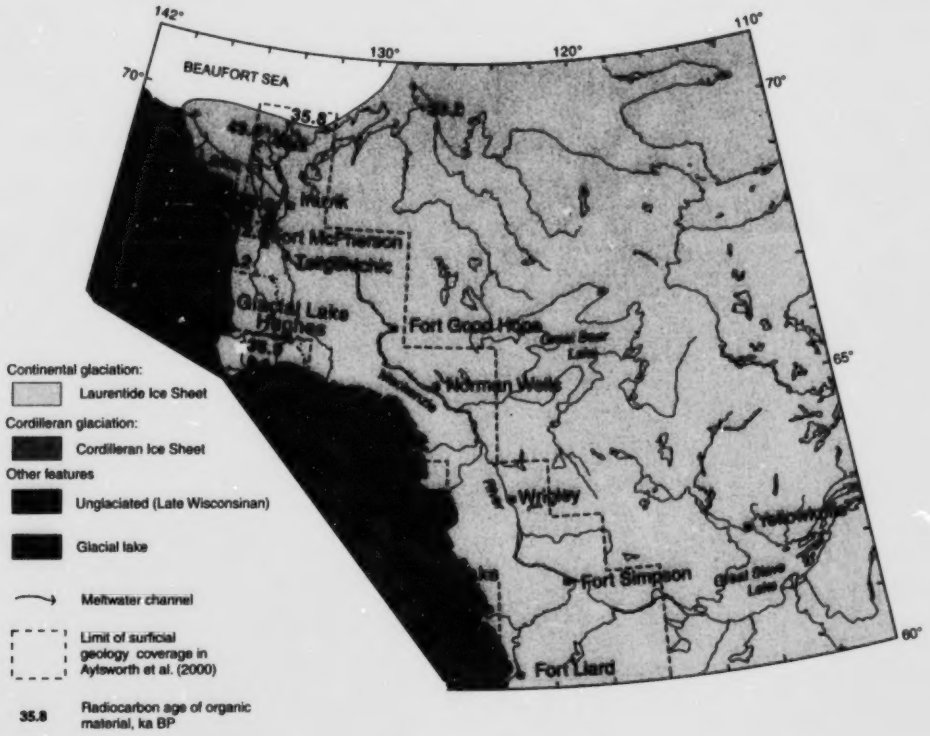

Figure 3. *Late Wisconsinan Laurentide maximum, 30–25 ka BP: glacial lakes Old Crow and Hughes at maximum draining via Porcupine River into Alaska; glacial Lake Nahanni forming.*

The Laurentide maximum (ca. 30 ka BP)

During the last glaciation, the Laurentide Ice Sheet moved west and northwest from the Canadian Shield and reached its maximum extent about 30 ka BP (Fig. 3). The ice sheet covered the Mackenzie and Peel plains and abutted the Mackenzie and Richardson mountains, with ice tongues extending up to 50 km up valleys such as those of the Keele, Mountain, Gayna, and Arctic Red rivers. In the Mackenzie Mountains, ice blocked pre-existing drainage and created a complex system of lakes and channels that drained to the northwest. This drainage system crossed the front of the Canyon Ranges and followed the edge of the ice sheet upon reaching Bonnet Plume Depression, where glacial Lake Hughes was formed by damming of the Peel River (Duk-Rodkin and Hughes, 1995). Glacial Lake Hughes spilled northward via the Eagle River outlet channel into glacial Lake Old Crow, formed in the Bell, Old Crow, and Bluefish basins by ice-sheet damming of the eastward-flowing Porcupine River. The outlet of glacial Lake Old Crow cut a canyon that permitted the water to drain westward into the Yukon River drainage system. At this time, montane glaciers formed only an ice cap on the Backbone Ranges of the Mackenzie Mountains.

Glacial lakes were also formed in most Mackenzie and Wernecke mountain drainages that were impounded by the ice sheet. The most important of these was glacial Lake Nahanni which occupied the South Nahanni and Flat river valleys. Thick lacustrine deposits (>30 m) occur in valleys along the southern boundary of the central Mackenzie Corridor (e.g. Mountain, Gayna, Arctic Red rivers).

Katherine Creek phase (ca. 22 ka BP)

By ca. 22 ka BP, the Laurentide Ice Sheet had retreated from its maximum position and readvanced to within less than 200 m of its maximum elevation (Fig. 4). This position records the limit of ice during the Katherine Creek phase of the Laurentide Ice Sheet, and is recorded by ice-marginal channels extending from the southern Mackenzie Mountains to the Richardson Mountains (Lemmen et al., 1994). A readvance is demonstrated in two localities where moraines of the Laurentide Ice Sheet truncate moraines of montane glaciers, which in turn crosscut the slightly higher Laurentide maximum. By this time, the Porcupine River had already been integrated as part of the Yukon River basin, and did not revert to its eastward drainage with the retreat of the Laurentide Ice Sheet. A fluctuating glacier margin in the Bonnet Plume Depression (Hughes et al., 1981) resulted in periodic

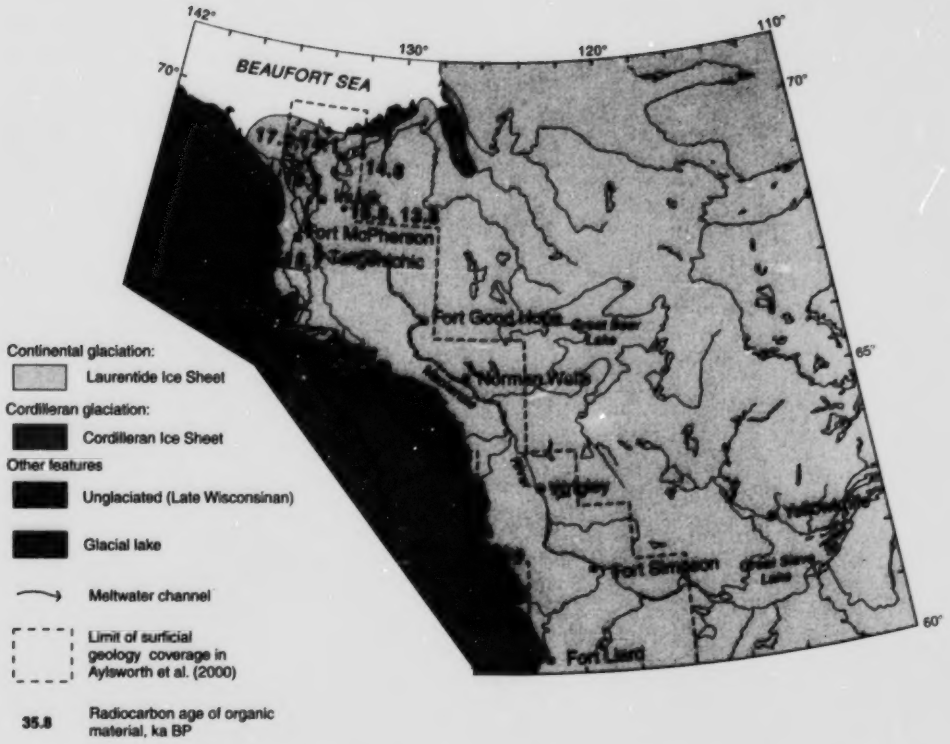

Figure 4. *Katherine Creek phase, 22–18 ka BP: Late Wisconsinan Cordilleran maximum; coalescence of Cordilleran ice with retreating Laurentide Ice Sheet; glacial Lake Nahanni at maximum.*

diversion of the Peel River drainage into Bell Basin. Montane glaciers in the Mackenzie Mountains (there were no glaciers in Richardson Mountains) had advanced to their maxima and contacted the Laurentide Ice Sheet in the Bonnet Plume, Arctic Red and Redstone river valleys. At the Arctic Red River, montane glaciers overrode stagnant Laurentide ice, producing loops of hummocky terrain (Duk-Rodkin and Hughes, 1991).

Glacial Lake Nahanni drained northward along the mountain front at this time, and remained dammed by the ice sheet until ca. 11 ka BP (Lemmen et al., 1994). Locally, the ice sheet impounded drainage, forming temporary glacial lakes on rolling moraine terrain along Peel plateau. Extensive ponding associated with this phase also occurred along the foothills between the Arctic Red River and the Bonnet Plume Depression.

Tutsieta Lake phase (ca. 13 ka BP)

The retreat of both the Laurentide Ice Sheet and montane glaciers occurred after ca. 22 ka BP (Fig. 5). The margin of the Laurentide Ice Sheet just prior to ca. 13 ka BP is recorded by the limit of the Tutsieta Lake phase (Hughes, 1987). At this time, meltwater drained freely along an ice-marginal channel, parts of which are occupied by the modern Hume, Ramparts, Arctic Red, Snake, and Peel rivers (Fig. 1). The Peel River incised the Peel plateau, draining parallel to the mountains and creating a new base level for the drainage of the Richardson Mountains between Bonnet Plume Basin and the Aklavik Range. Meltwater along the north and northeast sides of the ice sheet drained across the Anderson plains.

Local ice-damming of drainages formed temporary lakes, producing a discontinuous blanket of lake sediments that is most extensive on the rolling moraine of the Peel plateau. Postglacial stream incision of up to 500 m has produced spectacular canyons on the Peel plateau.

Glacial Lake Travaillant, Mackenzie River outlet, and Arctic Red River meltwater channel phase (ca. 12.5 ka BP)

Continued southward retreat of the Laurentide Ice Sheet and exposure of a drainage divide resulted in cessation of northward meltwater drainage across the Anderson plain at ca. 12.5 ka BP (Fig. 6). Instead, water flowed south towards the ice sheet and into glacial Lake Travaillant, which occupied the low area between Grandview Hills, and the Iroquois and Travaillant uplands. At its highest level, the lake spilled northward towards the Anderson plain, but later a westward outlet was

Figure 5. *Tutsieta Lake phase, 13 ka BP: retreat of Cordilleran Ice Sheet; drainage across Anderson plain; Cordilleran drainage exiting via Peel River.*

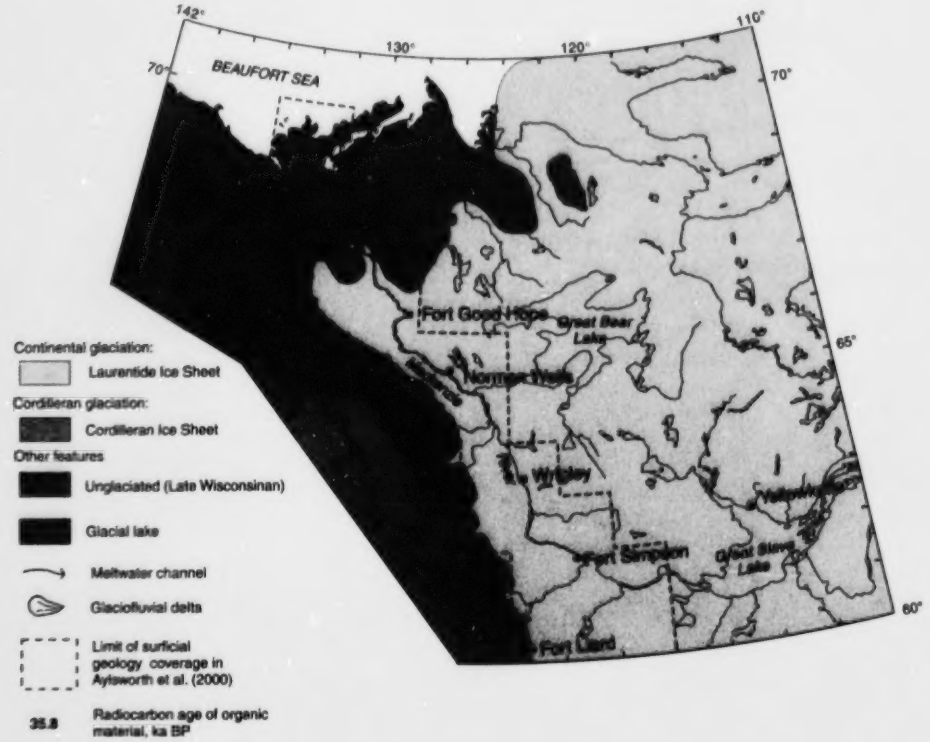

Figure 6. *Glacial Lake Travaillant phase, ca. 12.5 ka BP: formation of proglacial Lake Travaillant in Mackenzie valley; Cordilleran drainage exiting via Arctic Red River.*

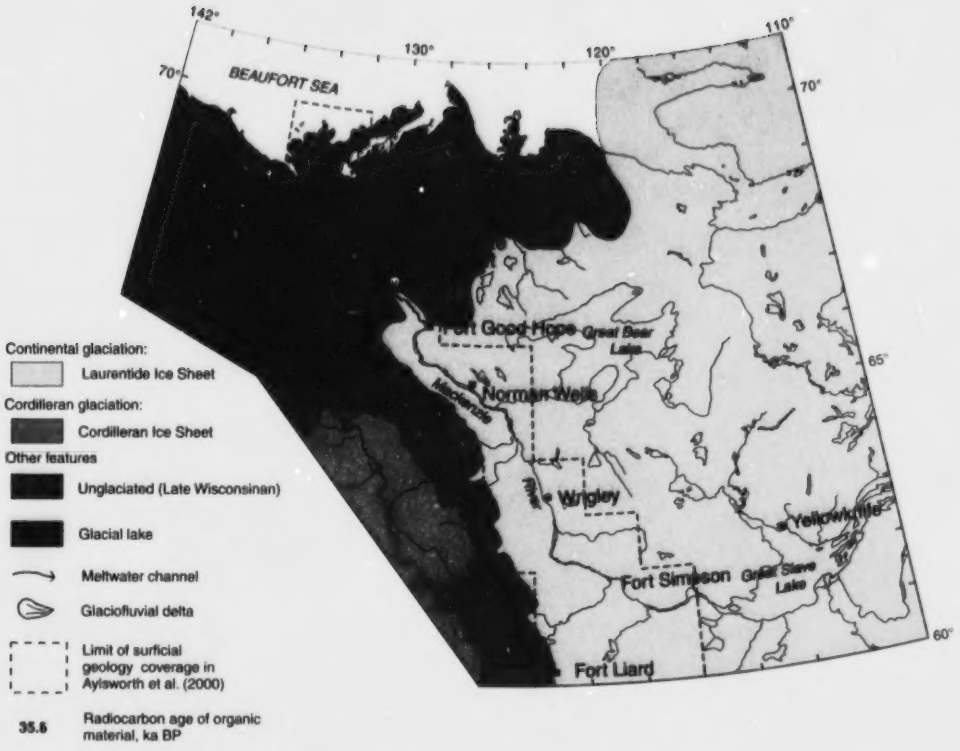

Figure 7. *Glacial Lake Ontaratue stage, ca. 12.2 ka BP: formation of proglacial Lake Ontaratue; draining of glacial Lake Travaillant. Cordilleran drainage exiting through Ramparts River.*

established via the Rengleng River, a few kilometres north of the modern Mackenzie River. With further ice retreat, the Mackenzie River became the outlet for glacial Lake Travaillant. An extensive glacial lake occupied upper Arctic Red River valley and drained via an ice-marginal channel that defines the modern route of lower Arctic Red River.

Both the Mackenzie and Arctic Red rivers cross extensive plains (approximately 3500 km^2) blanketed by glaciolacustrine sediments (3–30 m thick) deposited in glacial lakes during this stage.

Glacial Lake Ontaratue phase (12.2 ka BP)

Further retreat of the Laurentide Ice Sheet resulted in mountain waters flowing to the ice-sheet margin via the lower Ramparts River. The water was impounded against the slopes of a low interfluve forming glacial Lake Ontaratue. At its maximum level, the lake drained towards the Arctic Red River, but with continued ice retreat a second outlet (Ontaratue River) was established to the northeast with drainage entering the receding glacial Lake Travaillant (Fig. 7). By this time, the Mackenzie River had cut a canyon through bedrock at Tsiigehtchic (formerly Arctic Red River), establishing a permanent channel for meltwater originating from both sides of the ice sheet located upvalley.

Glacial lake sediments associated with this stage cover approximately 3000 km^2 with an average thickness of 15 m.

Early glacial Lake Mackenzie and Kelly Lake phase (12–11.5 ka BP)

Glacial Lake Ontaratue extended southward with retreat of the ice sheet, establishing the early stages of glacial Lake Mackenzie (Fig. 8). The northwest extension of glacial Lake Ontaratue was abandoned, with Ontaratue River serving as the lake outlet during surges of the retreating ice sheet (Duk-Rodkin and Hughes, 1995). The outlet of glacial Lake Mackenzie into the Mackenzie River was established at The Ramparts. At that time the ice sheet occupied the vicinity of Norman Wells around Kelly Lake and extended as far north as the Mountain River. The lower Mountain River served as the collector of meltwater drainage from the Mackenzie Mountains, and deposited an extensive deltaic complex into glacial Lake Mackenzie that extended from the mountain front to a few kilometres north of the confluence with the Mackenzie River. An ice-marginal channel along the slope of the Canyon Ranges was the major tributary to the Mountain River and carried waters from as far south as the Keele River.

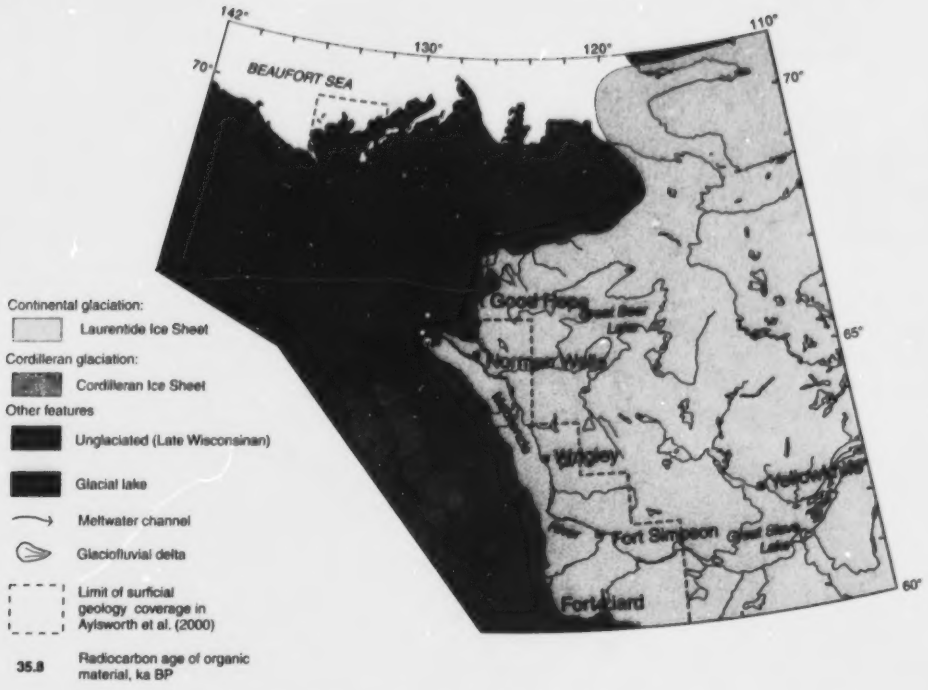

Figure 8. *Early glacial Lake Mackenzie, 12–11.5 ka BP: further downwasting of Cordilleran Ice Sheet, formation of glacial Lake Mackenzie; draining of glacial Lake Nahanni. Cordilleran drainage exiting via Mountain River.*

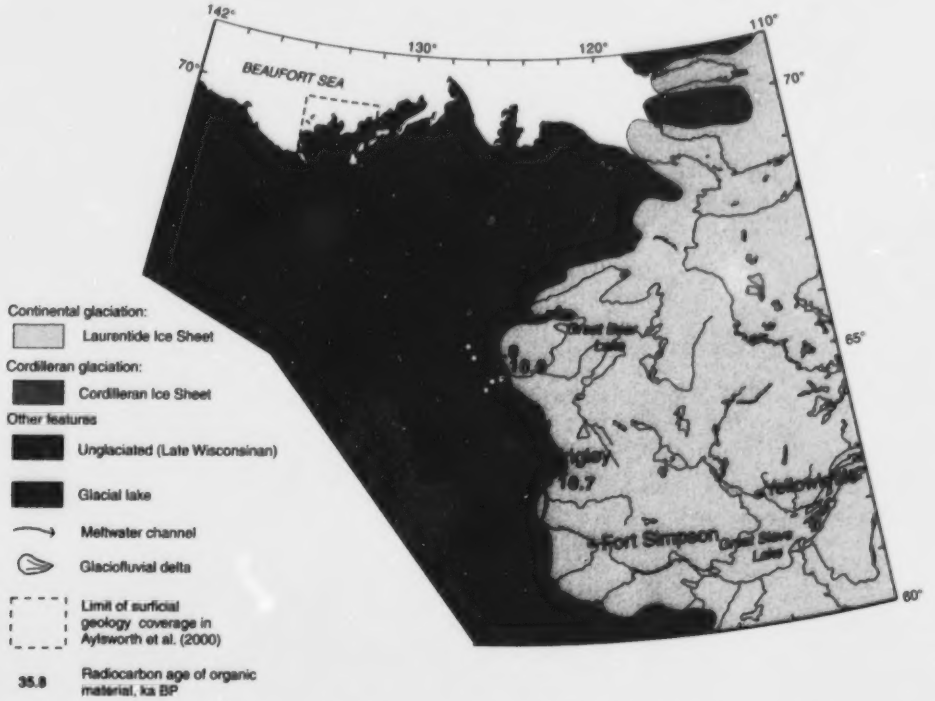

Figure 9. *Middle glacial Lake Mackenzie and early glacial Lake McConnell, 11.0 ka BP: glacial Lake Mackenzie occupying Mackenzie Lowland; early state of glacial Lake McConnell occupying Smith Arm of Great Bear Lake.*

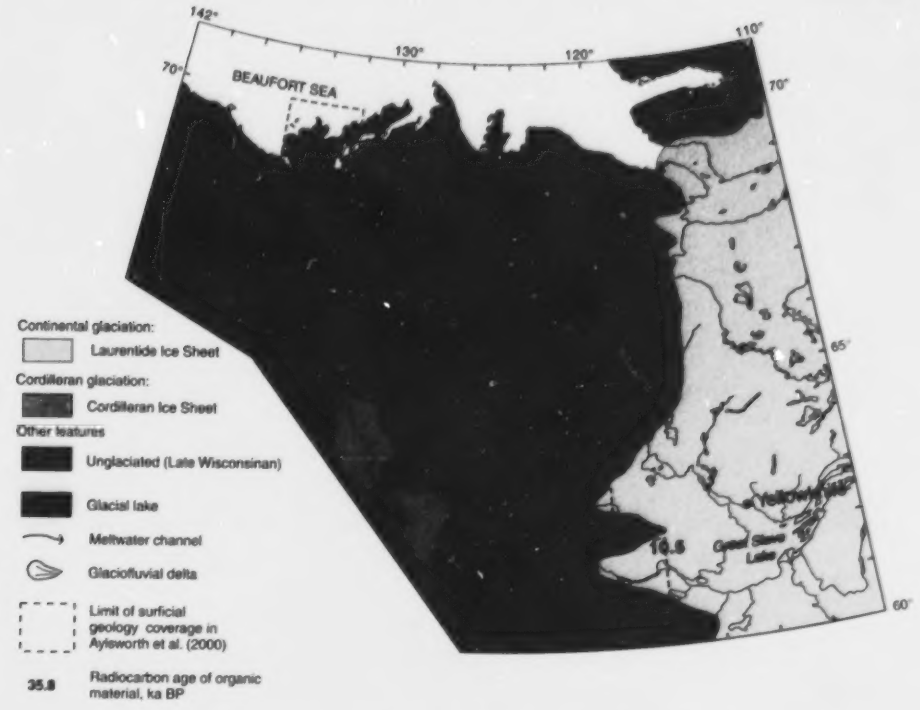

Figure 10. *Late glacial Lake Mackenzie and early glacial Lake McConnell 11–10.5 ka BP: glacial Lake Mackenzie at maximum; glacial Lake McConnell occupying Great Bear Depression; glacial Lake Liard draining into glacial Lake Mackenzie.*

Figure 11. *Glacial Lake McConnell, 10.0 ka BP: glacial Lake McConnell at maximum, occupying Great Bear and Great Slave basins.*

Glacial lake sediments extend along the Mackenzie River between the mouth of the Mountain River and Fort Good Hope (The Ramparts) and are up to 30 m thick.

Middle glacial Lake Mackenzie and early glacial Lake McConnell (11 ka BP)

Glacial Lake Mackenzie expanded southward, occupying the Mackenzie Lowland, in association with continued ice-sheet retreat (Smith, 1992). At ca. 11 ka BP, the ice sheet abutted the eastern and southern flanks of the McConnell Range and contacted the lake within the Mackenzie River valley (Lemmen et al., 1994; Fig. 9). Several meltwater channels contributed inflow to the eastern side of glacial Lake Mackenzie, constructing small deltas that indicate the lake reached a maximum elevation of 175 m a.s.l. In the southern Mackenzie Mountains, the Laurentide Ice Sheet still dammed glacial lakes in lower Liard and South Nahanni river valleys (Lemmen et al., 1994). Retreat of ice in the Great Bear Basin produced the earliest phase of glacial Lake McConnell, which drained via the Hare Indian River (Lemmen et al., 1994).

Lacustrine deposits related to this period form a discontinuous blanket along the lowlands of the Mackenzie River, which are more than 30 m thick in some locations south of the Great Bear River and north of the Redstone River.

Late glacial Lake Mackenzie and early glacial Lake McConnell (11–10.5 ka BP)

Retreat of the Laurentide Ice Sheet to the eastern flanks of the Horn Plateau by ca. 10.5 ka BP drained glacial lakes in the Liard and Nahanni river valleys, and resulted in the deposition of a large delta where Liard River entered glacial Lake Mackenzie (Lemmen et al., 1994; Fig. 10). At this time glacial Lake Mackenzie had expanded southeastward towards what would become the central basin of glacial Lake McConnell. Glacial Lake McConnell occupied an area slightly larger than present day Great Bear Lake, with its outlet via the route of Great Bear River which contributed to glacial Lake Mackenzie.

Glacial Lake McConnell (10 ka BP)

Although the Mackenzie Corridor was free of ice by 10 ka BP, it was still affected by Lake McConnell, which had expanded to become the second-largest Pleistocene lake in North America, covering all of the modern Great Bear, Great Slave, and Athabasca basins (Lemmen et al., 1994; Fig. 11). It has also been proposed that glacial Lake Agassiz, occupying much of Manitoba and significant areas of northern Saskatchewan and northern Ontario, flowed northward into the Mackenzie Basin, beginning about 9.9 ka BP (Smith and Fisher, 1993). If this hydrological connection existed, it would have approximately doubled the size of the Mackenzie River drainage basin for about 400 years.

ACKNOWLEDGMENTS

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Climate of the Mackenzie River valley

L.D. Dyke[1]

Dyke, L.D., 2000: Climate of the Mackenzie River valley; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 21–30.*

Abstract : The Mackenzie River valley is dominated by cold, dry air moving south from the high arctic in winter but receives a warm if brief summer as the arctic high pressure area retreats northward. Dry conditions are periodically modified by low-pressure cells moving from the Pacific Ocean or Beaufort Sea. Mean monthly temperatures in winter and summer tend to be uniform throughout the Mackenzie valley south of Inuvik. In winter, this is due to the parallelism between the arctic front and the Mackenzie valley as it traverses the region. In summer, longer daylight hours compensate for lower sun angles with increasing latitude to favour even heating. Snow and rainfall are low by North American standards but occasional heavy rains do occur. Throughout the 20th century, mean annual temperatures for most recording stations have on average risen between 1 and 2°C and snowdepth at the end of February has decreased.

Résumé : En hiver, la vallée du Mackenzie est envahie surtout par de l'air sec et froid en provenance de l'Arctique, alors qu'en été la migration vers le nord de la zone de haute pression de l'Arctique produit un climat chaud sur une courte période. Les conditions d'aridité sont périodiquement modifiées par des cellules de basse pression provenant de l'océan Pacifique ou de la mer de Beaufort. Les températures mensuelles moyennes en hiver et en été sont à peu près uniformes dans toute la vallée, au sud d'Inuvik. En hiver, l'uniformité des températures est attribuable au parallélisme de la vallée du Mackenzie et du front arctique au moment où celui-ci traverse la région. En été, les périodes d'ensoleillement plus longues compensent pour l'angle du soleil peu élevé sous cette latitude, ce qui favorise le réchauffement. Les précipitations nivales et pluviales sont faibles comparativement aux autres régions de l'Amérique du Nord, mais, à l'occasion, il y tombe des pluies abondantes. Tout au long du 20^{e} siècle, les températures annuelles moyennes à la plupart des stations d'enregistrement ont augmenté de 1 à 2 °C, et l'épaisseur du manteau nival de la fin de février a diminué.

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INTRODUCTION

The climate of the Mackenzie River valley, like the climate of any region, is fundamentally determined by the absorption of solar radiation by the Earth's surface and atmosphere. The strength of this radiation is determined by latitude, but snow cover, clouds, and the large change in hours of daylight (Fig. 1) and sun elevation during the year modify this input. Furthermore, the solar input is constantly redistributed between regions by atmospheric circulation. Thus, circulation causes air that has reached a thermal and moisture state elsewhere to arrive and influence the climate of the Mackenzie valley.

In winter, the Mackenzie valley is dominated by air originating in the polar region. The Cordillera protects this air from modification by easterly movement of mild, moist Pacific air and the very low sun angles throughout the Mackenzie valley in winter ensure an almost uniformly low solar input. In summer, the general circulation pattern changes markedly. Arctic air recedes, allowing low-pressure cells access from the west. Accompanying this change, air flow from the south and the long hours of sunlight act to make the area the warmest in all of Canada for its latitude (Stuart and Judge, 1991).

Climate is the primary factor determining the occurrence of permafrost and the intensity of landscape-forming processes. This paper describes the present-day climate of the Mackenzie valley, based mainly on climate normals for the interval 1961–1990 (Environment Canada, 1993). In this sense, this description is somewhat updated from the only thorough treatment of Mackenzie valley climate in existence, the two-volume monograph by Burns (1973). The almost complete limitation of climate recording stations to communities along the generally linear transect formed by the Mackenzie River and Delta restrict the accuracy with which parameters such as temperature or snowfall can be depicted by conventional contour maps. Although this paper makes use of maps for describing the behaviour of the atmosphere in this region, it otherwise adopts a corridor approach to depicting climate parameters along the Mackenzie valley.

The mean annual solar radiation reaching the area is shown in Figure 2 (Hare and Thomas, 1979). This is the amount of heat potentially available for warming the ground and the air near the ground. It is dependent on latitude and the amount of radiation returned to space by reflection from cloud cover. Much of the radiation received at the ground surface is also returned to space due to reflection from the ground surface, and later by re-radiation from the ground. This returned amount is roughly equal over the region. When the re-radiated amount is subtracted from the incoming amount, net radiation, or the amount left to warm the ground, remains. Net radiation has a proportionately much larger decrease from south to north than mean annual radiation. This indicates the gradation of net solar heat input present along the Mackenzie valley.

TEMPERATURE

The mean monthly temperature for the Mackenzie valley is shown in Figure 3. This figure (and other similar ones) treat the meteorological stations located along the Mackenzie River out to the Beaufort Sea at Tuktoyaktuk as a north-south transect. This way of depicting a climatic parameter makes variation along the axis of the valley readily apparent; however, it must also be realized that all stations are in valley-bottom locations and no information on the relationship of temperature to elevation is contained. An interesting feature of the mid-summer and mid-winter temperature distribution in the Mackenzie valley is its general uniformity south of Inuvik. The highest and lowest mean monthly temperatures vary by no more than 5°C throughout this section.

Atmospheric temperature inversions, whereby the expected decrease in air temperature with altitude is reversed, are common throughout the Mackenzie valley, especially in winter (Stuart et al., 1991). Although detailed records of temperature changes with altitude are available for Fort Smith, Norman Wells, and Inuvik, there is no corresponding information for temperature change with increasing elevation. Inversions in winter at Norman Wells typically result in air temperatures at an altitude of 1000 m which are 10°C warmer than at ground level. This is approximately the elevation of the top of the mountain ridge immediately east of Norman Wells (Norman Range of the Franklin Mountains). Measurements to determine the correspondence of temperature with elevation on this ridge are underway by the Geological Survey of Canada (Taylor et al., 1998).

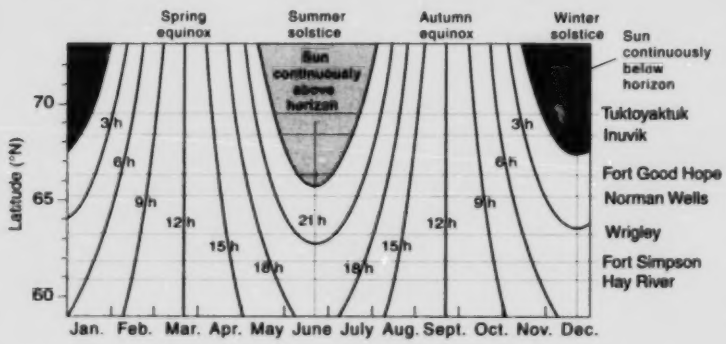

Figure 1.

Hours during which the top of the sun is above the horizon for any latitude in the Mackenzie River valley at any time during the year. The sun never sets in summer for at least a short time north of latitude 65°43'N, increasing to almost two months for the latitude of Tuktoyaktuk. Similarly, the sun never rises in the winter for at least a short time north of latitude 67°23' and Tuktoyaktuk is continuously without direct sunlight for almost two months. The long hours of sunlight in summer and long darkness in winter contribute to the uniformity of high and low temperatures throughout the region.

Differences in the overall amount of freezing or thawing air temperature throughout the Mackenzie valley show more contrast than the maximum and minimum monthly temperatures, as shown by the transect of freezing and thawing indices in Figure 4. Although the information shown in Figures 3 and 4 is essentially the same, freezing and thawing indices express, in one number, the combined effect of duration and magnitude of freezing and thawing temperatures. They are a commonly available, useful alternative to radiative heat-flux measurements for estimating ground temperature. The considerable increase in proportion of freezing degree days northward along the valley reflects the increasing dominance of winter high-pressure systems which travel from the high Arctic generally southeasterly (Fig. 5). By January, an extensive ridge of high pressure is established over the Mackenzie valley, forming an effective block to the westward movement of warm, low-pressure areas which tend to reside in the Gulf of Alaska (Fig. 6). This dominance continues from December through March (Environment Canada, 1988). The uniform cold temperatures pointed out in Figure 4 result, in part, from the almost simultaneous incursion of the arctic air mass over all parts of the Mackenzie valley in the winter, as indicated by the residence times of arctic air in Figure 7. Warm intervals during the winter are caused by warm air masses that are part of low-pressure cells that displace the Mackenzie valley high-pressure ridge.

PRECIPITATION

Precipitation is restricted in the Mackenzie valley because of the rain-shadow effect of the Cordillera. The valley is typified by a precipitation of 300 to 400 mm, tapering off along the Mackenzie Delta to 250 mm at Inuvik and 125 mm at Tuktoyaktuk. Low-pressure systems enter the Mackenzie valley from several sources, but the channelling effect of the valley results in two major routes (Fig. 8). The first involves air which originates in the northern Pacific Ocean and moves through Alaskan valleys into Yukon Territory and then into the Mackenzie valley, often moving south from there. Air that moves directly south from the Beaufort Sea combines with

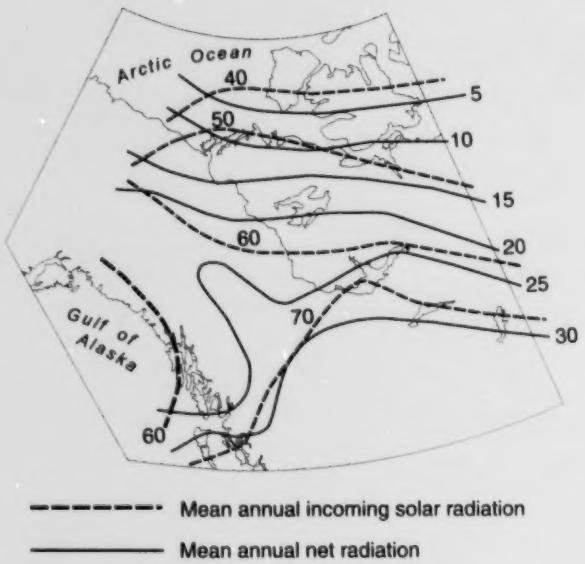

Figure 2.

Annual incoming solar radiation (kilolangleys = kilocalories/cm^2) is the basis for the Mackenzie valley's thermal character (dashed lines). Radiation input decreases with latitude, but is also reduced by cloud cover. Annual net radiation (solid lines) is the amount of energy that reaches the ground and remains to warm it. Although net radiation is ultimately responsible for determining air temperature, heat is redistributed by the movement of air masses to give a climate different from that determined by radiation at a particular location. In the case of the Mackenzie valley, heat is removed by cold air from the north, but brought in by warm air from the west or south (map adapted from *Hare and Thomas, 1979).*

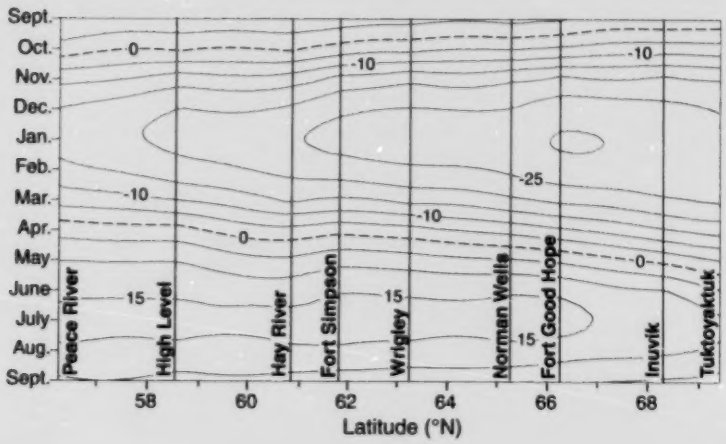

Figure 3.

Mean monthly temperature (°C) throughout the year, based on the 1961–1990 climate normals (Environment Canada, 1993) for settlements along the Mackenzie valley. The observations have been contoured to show the general trend of temperatures spatially. Note the generally similar mid-summer and mid-winter mean monthly temperatures for the valley south of Inuvik.

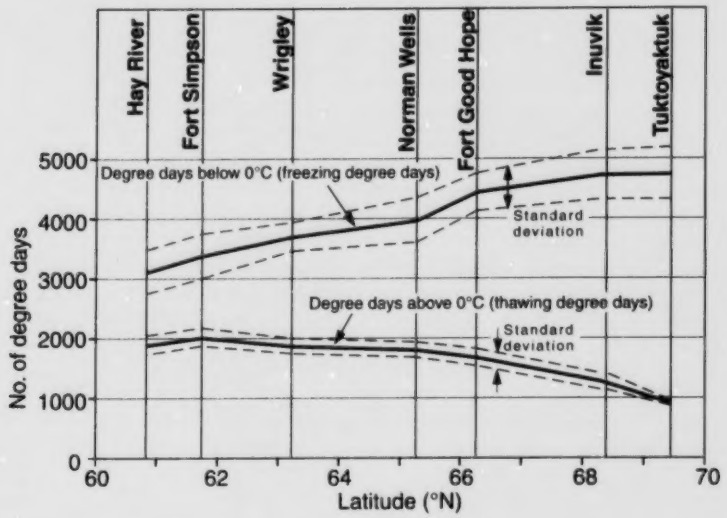

Figure 4.

A transect of freezing and thawing indexes (degree days above and below 0°C) along the Mackenzie valley. The freezing index must usually be greater than the thawing index for permafrost to persist. This is due mainly to the insulating effect of snow cover and the consequent reduction in the freezing index at the ground surface; however, in peatlands, this effect can be offset by the insulating effect of dry peat in summer. Under this condition, permafrost can persist when the indices are roughly equal. The wider standard-deviation band for the freezing index is due to the greater winter temperature variability associated with influxes of warm air from the Pacific Ocean.

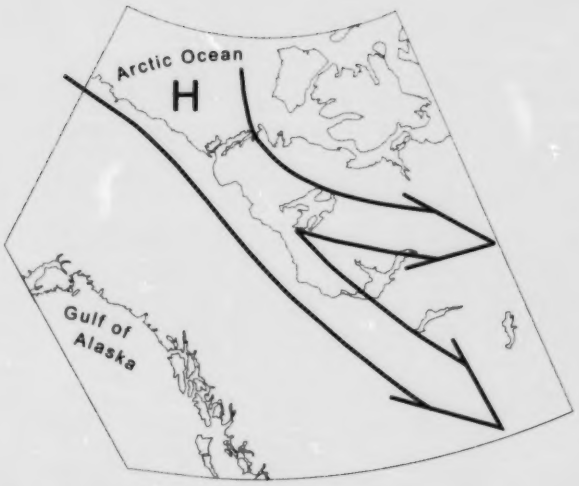

Figure 5. *Prevailing routes for the movement of arctic air into the Mackenzie valley during November to March. This air proceeds down the valley and continues to the east or southeast to act as the source for cold air masses that reach southern Canada and the United States* (adapted from *Burns, 1973*).

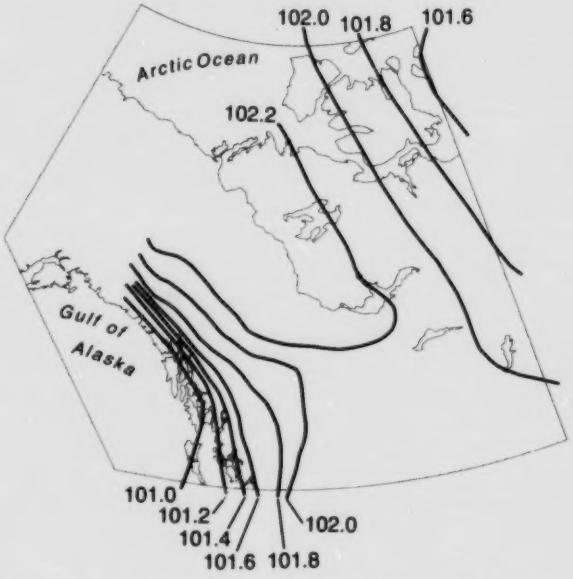

Figure 6. *Mean sea-level atmospheric pressure (kPa) for January, based on the 1951–1980 climate normals. This pattern is typical of the winter, and features the dominating ridge of high pressure centred over the Mackenzie valley. The steep gradient to low pressure in the Gulf of Alaska results from low-pressure cells that typically reside in this area. Occasionally, they displace the high-pressure ridge and move eastward across the Mackenzie valley* (adapted from *Environment Canada, 1988*).

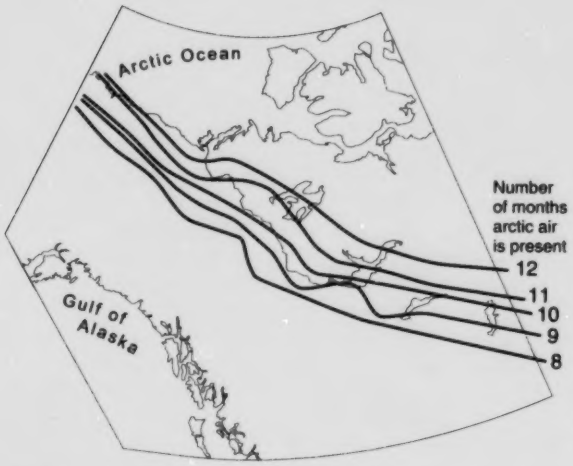

Figure 7. *Dominance of arctic air in the Mackenzie valley region. The contours represent the number of months per year arctic air is present over any location. The front shifts position during the year, remaining north of the Mackenzie valley during the summer, then moving south of it in late fall. The parallelism between this moving boundary and the valley contributes to the uniform cold temperatures in winter (diagram* adapted from *Bryson and Hare, 1974).*

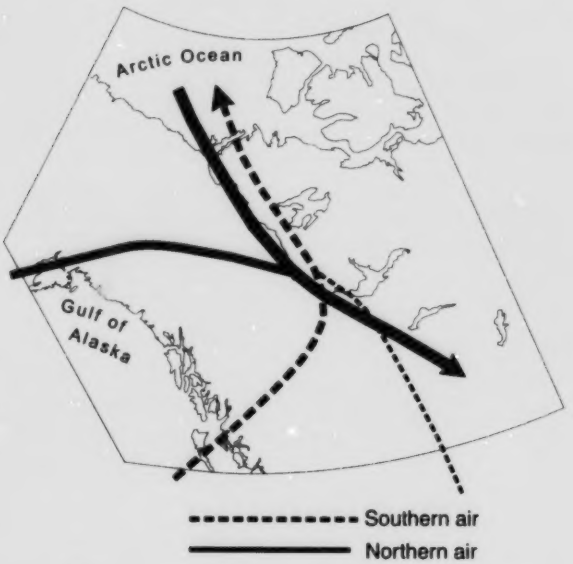

Figure 8. *Major routes of low-pressure systems entering the Mackenzie River valley. Decreasing line prominence indicates decreasing frequency with which paths are followed. Routes from the Beaufort Sea and Gulf of Alaska are the primary paths for low-pressure systems in summer and fall. Systems moving from the south (dashed lines) are infrequent, but are most likely to bring the heaviest precipitation, as was the case with the 1988 record heavy rains in the southern valley* (adapted from *Burns, 1973).*

this flow to form the primary air pathway. The second pathway draws air from further south in the Pacific Ocean along routes that break through the Cordillera from central British Columbia to as far south as the Columbia River. This air then moves north into the Mackenzie valley by way of the Liard River valley or the plains to the east of the Cordillera.

Individual daily rainfalls are typically light with few exceeding 5 mm (Fig. 9); however, heavy rains can happen, as indicated by the summer occurrence of daily rainfalls occasionally exceeding 50 mm for sites south of Inuvik. Most of these heavy rains result from local convective storms and are inadequately represented by records from climate stations. Observations of local flooding, such as the 1970 flood of the Keele River (Burns, 1973), are evidence of otherwise unrecorded heavy rains. Also, severe frontal storms occasionally reach the region. For example, two storms moving into the Mackenzie valley from late June to mid-July, 1988, along the least-frequently traversed route shown in Figure 8, resulted in the highest recorded rainfalls for the entire region (Jasper and Kerr, 1992). Between June 29 and July 3, 260 mm of rain fell at Nahanni Butte. These storms produced a maximum discharge of 33 700 m^3/s for the Mackenzie River at Norman Wells, the highest discharge ever measured there.

By November, most precipitation falls as snow. Mean monthly snowfall rises sharply in the fall and then diminishes (Fig. 10) as the high-pressure area develops in the Mackenzie valley and prevents humid air moving in from the west. Snow accumulation proceeds more steadily throughout the fall and winter (Fig. 11), as the decrease in snowfall is compensated for by a decrease in the likelihood of thaws. The snowpack reaches its maximum depth by some time in March and then begins a more rapid ablation as the high-pressure area breaks up. As with rainfall, snowfall at Tuktoyaktuk is low compared with stations further up the Mackenzie valley, although it is typical of arctic coastal values. This is due to the dominance of arctic air, and relative remoteness from moisture-carrying storm tracks further south.

WIND

Wind patterns in the Mackenzie valley are complex due to the mountainous nature of the region. Local influences include chinook winds from the Cordillera, channelled flows through major valleys such as the Liard valley, and diurnal winds resulting from surface-temperature differences. Through the bulk of the Mackenzie valley, there is a general northwest-southeast trend to the wind direction due to the channelling effect of the valley. Of the record sites available, this trend is seen most markedly at Norman Wells and Fort Simpson (Fig. 12). Winds at Inuvik and Tuktoyaktuk show a prominent component from the east, reflecting circulation in the arctic high-pressure area, unhindered by the low surroundings. Differences in temperature between water and land offset other influences at Hay River, resulting in no strongly preferred direction.

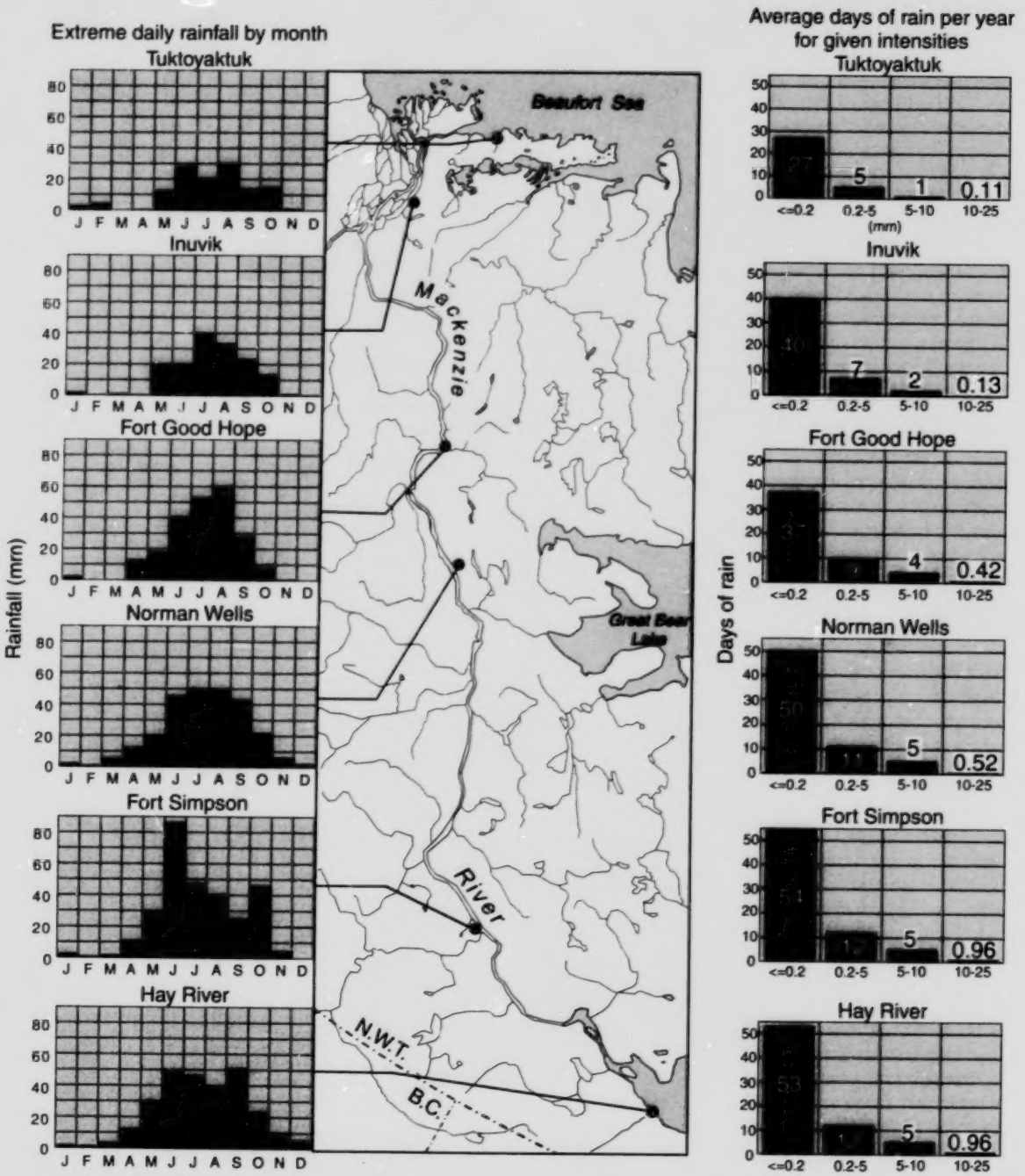

Figure 9. *Rainfall intensity along the Mackenzie valley. The bar graphs on the right show the average number of days per year having rainfall of given amounts. The graphs on the left show the single most extreme rainfall for each month during the interval 1961–1990. Rainfalls tend to be most frequent and heaviest in the southern half of the Mackenzie valley because of the incursion of moisture-laden North Pacific air masses. Furthermore, Pacific air is most likely to be unstable and produce the convective storms that are usually associated with heavy rains (diagram compiled from 1961–1990 climate normals (Environment Canada, 1993)).*

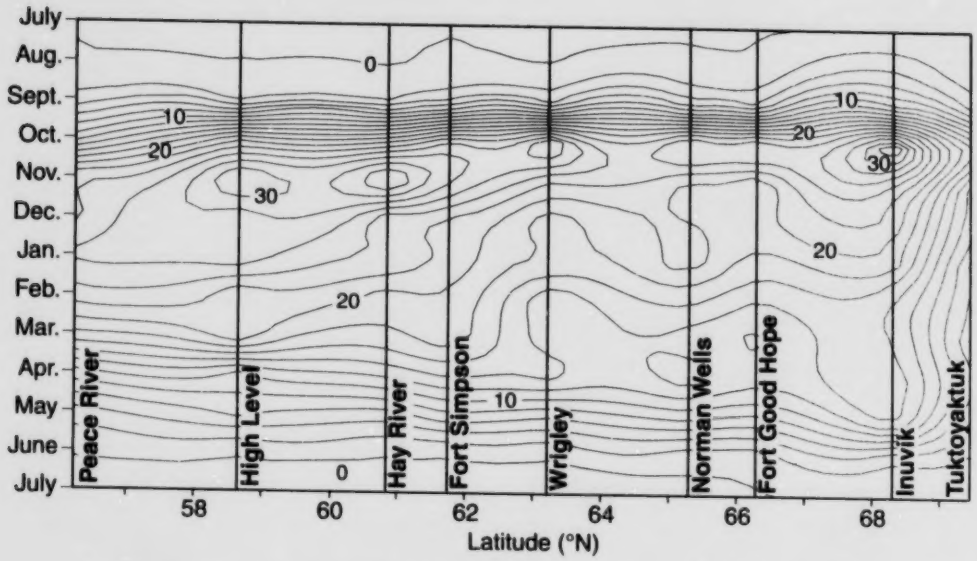

Figure 10. *Mean monthly snowfall (in centimetres) generalized along the Mackenzie valley with a continuation into northern Alberta. Snowfall is heaviest in the fall, before the arctic high-pressure ridge, pictured in Figure 6, has established itself and formed a barrier to the eastward movement of low-pressure cells. By the time this high dissipates in the spring, warming reduces the chance of precipitation as snow (diagram compiled from 1961–1990 climate normals (Environment Canada, 1993)).*

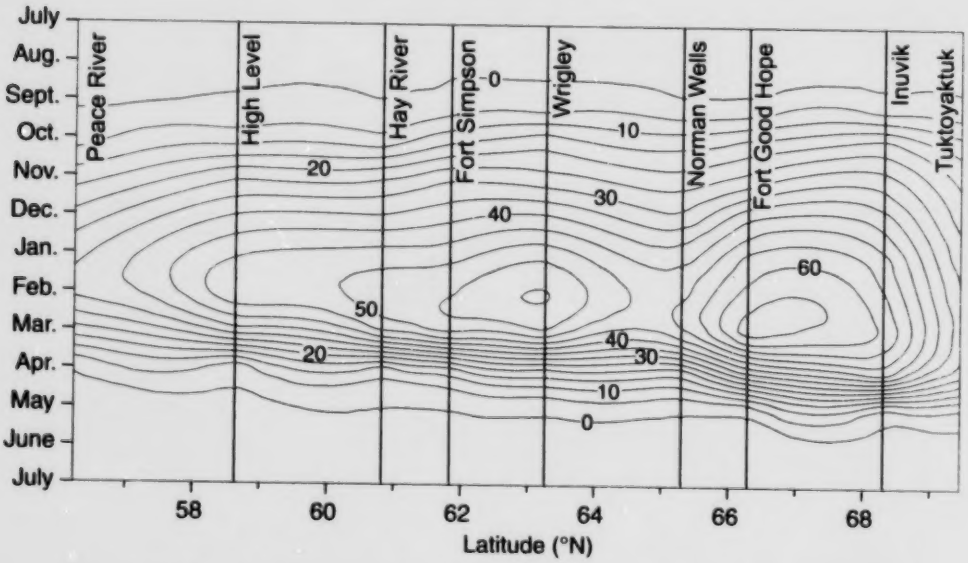

Figure 11. *Snow depth at month's end (in centimetres) generalized for a transect following the Mackenzie valley. Heavier snowfalls in fall are still subject to thawing episodes, whereas lighter falls in winter seldom thaw, resulting in an overall steady accumulation. Decreasing snowfalls in warming temperatures in spring lead to a rapid decrease in snow depth.*

RECORDED CHANGES

This summary is based on averages of measurements over a fixed interval of time, and so gives no indication of change. For some stations, climatic records exist prior to the 30 year interval on which much of this summary is based. Measurement records are thus available that can be used to discern change over intervals occasionally longer than 50 years. For several Mackenzie valley stations, plots of average annual temperature and snow depth at the end of February are presented in Figure 13. For both these parameters, linearregression lines have been included in the histories. The regression lines show the overall average trend for the period of record. Although short-term variability is greater than the regression trends for both parameters, all locations show an overall increasing temperature and decreasing snow depth for roughly the last 50 years, except Fort Good Hope, where a slight decrease in temperature has occurred. Although ground temperatures would be expected to respond to the overall warming observed in Figure 13, Stuart et al. (1991) have noted that the corresponding decrease in snow depths may offset this

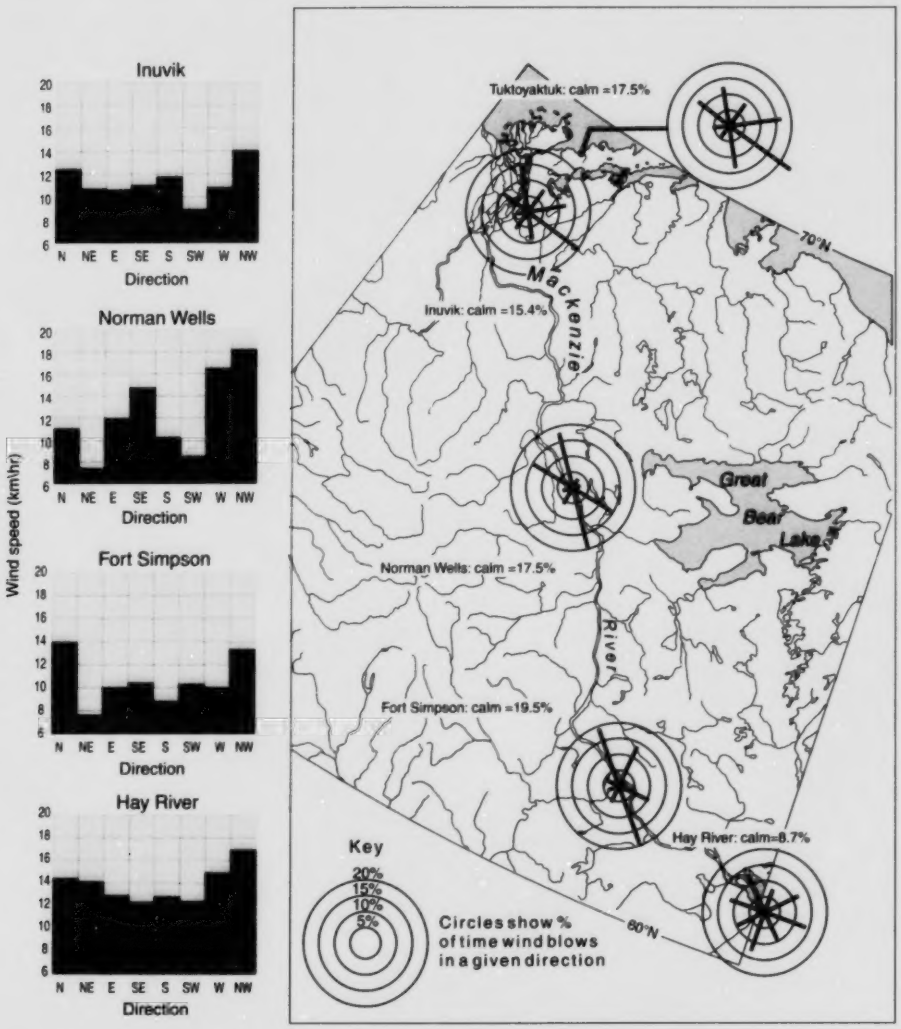

Figure 12. *Wind direction and speed for settlements in the Mackenzie valley. Winds tend to be channelled by the Mackenzie valley, being northwesterly for outbreaks of the arctic high and southeasterly when this air is replaced by warmer air from the south. Dominant easterlies on or near the arctic coast result from the clockwise circulation about the arctic high. Winds at Hay River suggest the influence of onshore-offshore temperature differences beside Great Slave Lake competing with more regional patterns of flow. Average wind speeds are highly variable, but show a slight tendency to be strongest from the north or northwest and the southeast, again reflecting the channelling effect of the valley.*

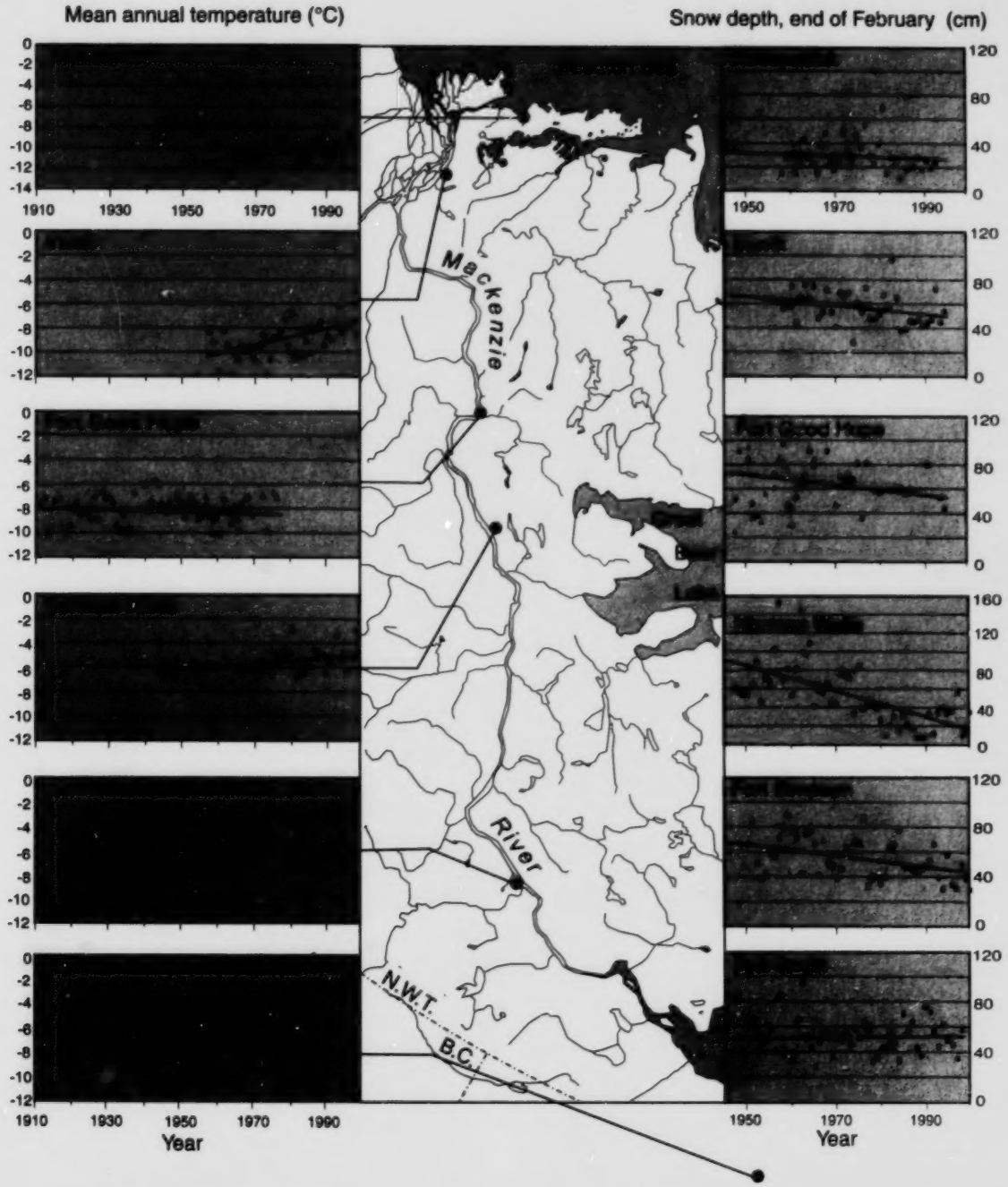

Figure 13. *The six longest Environment Canada climate records available for mean annual temperature and snow depth at the end of February for stations in the Mackenzie valley. Linear regression lines suggest an overall warming and decrease in snow depth for the intervals of record. These trends are consistent with other indications of climate change during the emergence from the Little Ice Age, such as retreat of glaciers, ground temperature warming, and degradation of permafrost near its southern limit.*

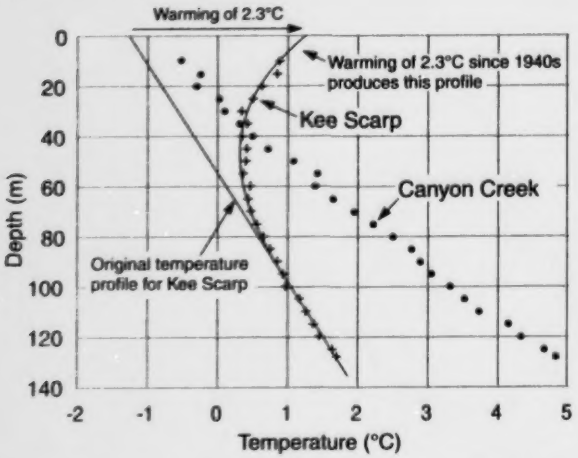

Figure 14. *Ground temperatures measured in the Kee Scarp borehole (crosses) and Canyon Creek borehole (dots) near Norman Wells. The deepest values for the Kee Scarp borehole are used to define the initial linear temperature gradient that is assumed to have existed before a warming in air temperature produced the curvature in the observations. The difference in temperature given by the two trends at the surface is the inferred warming. The Canyon Creek borehole shows no curvature, but projects to the same surface temperature as probably originally existed at Kee Scarp. Atmospheric temperature inversions in winter may be delaying the response to an overall warming that is apparent in the higher elevation Kee Scarp borehole.*

change by providing less insulation against cold penetration in winter. Therefore, ground warming may be delayed by decreasing snow depths.

USE OF GROUND TEMPERATURES FOR INFERRING PAST AIR TEMPERATURES

Independent evidence indicating an air temperature rise at Norman Wells is available from ground-temperature measurements in a 128 m borehole on Kee Scarp, a limestone ridge lying about 300 m higher than, and a few kilometres from, the Norman Wells climate station (Burgess et al., 1990). The temperature change with depth exhibits strong curvature that can be attributed to climate warming during the last few decades (Fig. 14). A change in temperature at the ground surface has been propagated downwards, displacing temperatures from the original linear gradient. The time since the beginning of the change can be determined by matching predicted temperature profiles with the borehole data. The analysis of the Kee Scarp borehole shows that the curvature in the temperature profile is consistent with a gradual ground-surface temperature increase of 2.3°C starting in the mid-1940s or a sudden increase in the mid-1960s (A.E. Taylor, pers. comm., 1996).

The inferred temperature increase at the Kee Scarp site is greater than the average annual air-temperature warming of 1°C over the same time interval indicated for Norman Wells (Fig. 13). A forest fire which burned the area around the Kee Scarp borehole, and probably destroyed insulating surface vegetation in 1954, may account for part of the warming. In contrast, a temperature profile to the same depth at Canyon Creek, another location near Norman Wells, but at the same elevation as the climate station, shows no suggestion of a temperature change (Fig. 14). The equilibrium temperature profiles for both boreholes give the same temperature when projected to the ground surface, differing in slope because of different ground thermal conductivities at the two sites. The higher elevation of the Kee Scarp site may be associated with a more pronounced air-temperature warming because of the tendency for the coldest air to pool at the lowest elevations during inversions. The lack of a temperature change at the Canyon Creek site may be due to a compensating effect of decreasing snow depth if the trend of decreasing snowfall at Norman Wells indicated in Figure 13 is representative of the area. Although these explanations for the ground temperatures at the two sites cannot be verified, the difference in temperature between the two boreholes is a suggestion of how important local climate may be in controlling ground temperature and the distribution of permafrost.

ACKNOWLEDGMENTS

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Permafrost distribution and ground ice in surficial materials

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Heginbottom, J.A., 2000: Permafrost distribution and ground ice in surficial materials; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 31–39.*

Abstract: The Mackenzie valley lies within the permafrost region of Canada; in consequence, most moisture in the ground occurs as ground ice. Ground ice commonly forms a significant component of the upper part of the ground and has important implications for economic development in the region. Ice occurs in many forms, most often as fillings in the intergranular pores of soils; however, it can also form much more massive bodies such as ice wedges and layers up to several metres thick.

Climate warming will lead to thawing of ground ice. This will lead to surface subsidence, increases in slope failures, and silt deposition in water bodies. Wildfire reduces the insulating capacity of ground-surface vegetation and will also induce excessive thaw. These outcomes can affect foundations and the stability of structures. Any cooling of the climate would reverse most of these effects, and would have relatively few adverse impacts.

Résumé : La vallée du Mackenzie est située dans la région pergélisolée du Canada. En conséquence, la majeure partie de l'humidité du sol se présente sous forme de glace de sol. La glace de sol constitue généralement une composante importante de la partie supérieure du sol et a, de ce fait, des répercussions de taille sur le développement économique de la région. La forme de la glace est variée, mais le plus souvent, elle remplit les espaces interstitiels. Elle peut également former des amas plus vastes comme des coins de glace et des couches de plusieurs mètres d'épaisseur.

Le réchauffement climatique causera le dégel de la glace de sol. Il s'ensuivra des subsidences de la surface, des glissements de terrain plus nombreux et une sédimentation de silt dans les étendues d'eau. Les feux de forêt, en éliminant la végétation de surface, réduisent l'isolation de la surface et provoquent un dégel excessif, ce qui peut avoir des effets néfastes sur les fondations et la stabilité des structures. Tout refroidissement du climat inverserait la plupart de ces effets et aurait relativement peu de répercussions de sens opposé.

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INTRODUCTION

The Mackenzie valley lies entirely within the permafrost region of northwestern Canada. Within this region, the temperature of the ground is continuously below 0°C over significant proportions of the area. One result of this is that moisture in the ground is commonly frozen, with ice often forming a substantial proportion of the upper part of the ground. The ground surface throughout the area is also subjected annually to deep seasonal thawing and freezing in response to seasonal fluctuations of air temperature. Both the low ground temperatures and the presence of ground ice have major effects on the landscape, on active geomorphic processes, and on construction and development in this region.

The ground ice occurs in two main forms, as structure-forming ice, bonding the enclosing sediments, and, particularly in the northern part of the study area, as large bodies of more or less pure ice. These two groups are discussed separately. The structure-forming ice comprises segregated ice, intrusive ice, reticulate vein ice, ice crystals, and icy coatings on soil particles. The large bodies of more or less pure ice, which exist mainly in the upper part of the ground, occur as pingo cores, massive icy beds, and ice wedges.

The distinction between structure-forming ice and bodies of massive ground ice is important for engineering and construction purposes. The design, construction, and operational plans for engineering works need to accommodate structure-forming ice beneath the foundations, as the presence of such ice is ubiquitous. The larger ground-ice bodies can often be avoided, however, provided their presence can be identified and their location detected. Landform analysis and geophysical exploration techniques can be of great value in addressing this problem.

For nearly all economic or development activity in northern regions, the occurrence of ground ice within the permafrost is of much greater significance than the temperature of the ground. This is because of the ground-stability problems associated with any surface disturbance that may cause subsequent thawing of ice-rich permafrost. Thus sufficient and accurate information on the character, distribution, and form of frozen ground and ground ice, as well as the geographical and geological setting of its existence, is very important for rational planning of economic development in northern Canada. In various ways, permafrost has always had significant effects on the economic development of the North, particularly for the energy and mining industries, but also for the construction of modern settlements and infrastructure elements such as roads, railways, airfields, and utilities.

The details of the permafrost extent and ground-ice conditions shown in the accompanying figures (Fig. 1a, 1b) have been summarized largely from engineering geology field mapping by the Geological Survey of Canada, as compiled by Heginbottom and Radburn (1992) and Heginbottom and Ruhland (unpub. map manuscript, 1995). Where data from engineering geology field maps were lacking, maps of surficial geology and Quaternary features were relied on. The compilations by Heginbottom and Radburn (1992) and Heginbottom and Ruhland (unpub. map manuscript, 1995) were produced by evaluating all the permafrost- and ground-ice-related information provided by the field geologists and map compilers, developing a common legend for the permafrost and ground-ice conditions, and drafting the maps with a minimum of generalization (Heginbottom, 1983, 1985). The source maps for these compilations were published between 1966 and 1982 — a time span of 17 years. This imposes some limitations on the data for the purposes of evaluating the implications of climate change, in that the climate was not regarded as changing over the period from 1966, when the first source map was published, until the present — a situation that is now known to be unrealistic. Thus the maps by Heginbottom and Radburn (1992) and Heginbottom and Ruhland (unpub. map manuscript, 1995) present a summary of the permafrost and ground-ice conditions as reported, but the data have not been adjusted to a common date.

The Quaternary geology of the region has been described by Hughes (1972), Rampton (1982, 1988), Hughes et al. (1983), Vincent (1989), and Duk-Rodkin and Hughes (1995). Ground ice in the Arctic Lowlands in general was reviewed by Carter et al. (1987). Terminology in this paper generally follows the recommendations of the Permafrost Subcommittee (1988), except for terms and definitions of permafrost zones; these follow Heginbottom and Radburn (1992).

EXTENT OF PERMAFROST

The extent of permafrost increases from sporadic discontinuous permafrost (with <35% of the area underlain by frozen ground), in the southern part of the region, to continuous permafrost (>90% of the area underlain by frozen ground), at the Beaufort Sea coast (Fig. 1b). There are two exceptions to this broad generalization. First, the Mackenzie Delta comprises an area of intermediate discontinuous permafrost (35–65% of the area underlain by frozen ground), well within the zone of continuous permafrost. This is due partly to the large proportion of the delta occupied by lakes with subjacent taliks, but also to warming of the ground by advected heat from the Mackenzie River. Secondly, in the Franklin Mountains, the Canyon Ranges of the northern Mackenzie Mountains, and the Richardson Mountains, altitudinal effects result in more extensive permafrost than would otherwise be expected on the basis of latitude alone. These effects occur both at upper elevations, as a result of normal adiabatic temperature-lapse rates, and also in valley bottoms, caused by ponding of very cold air and severe radiative cooling under conditions of stable temperature inversions in winter.

GROUND-ICE CONTENT OF SURFACE DEPOSITS

The ice content of the upper part of the ground ranges from negligible in consolidated rock and dry, coarse-grained deposits in the southern part of the study area, to high in peatlands and fine-grained deposits, particularly in the central and northern parts of the study area (Fig. 1a). The ice usually occurs as frozen pore water, lenses of segregated ice

typically a few millimetres thick, or ice veins. These ice forms, which act to bind the enclosing sediment, lead to the concept of 'ice-bonding'. This provides a practical distinction from the strictly thermal condition of being below 0°C used to define 'permafrost'.

At most locations, the upper few metres of permafrost, immediately below the base of the summer thaw zone (termed the active layer), tend to have a relatively higher ice content than deeper permafrost (Pollard and French, 1980). This has been postulated to have resulted from a general cooling of the climate over recent decades to centuries. The ice-content figure (Fig. 1a) summarizes current knowledge of the general content of ice in the upper 10–20 m of the ground, in six categories.

High ice-content permafrost

High ice-content permafrost comprises ground containing greater than 15% visible ice. Areally, this is the least extensive of all the categories, being essentially confined to an area of peat-covered, glaciolacustrine silts and clays, immediately northwest of Sans Sault Rapids, and to a few small areas of fine-grained deposits to the west of the Richardson Mountains, and in the upper valleys of the Peel and Arctic Red rivers.

Moderate ice-content permafrost

Moderate ice-content ground contains between 5 and 15% visible ground ice in the upper part of the ground. This is the most widespread category of ground ice in the study area, and occurs throughout the region from 60°N to the Arctic coast. It is the most commonly occurring form of ground ice in areas of till plains, including those with extensive peat cover, and the Mackenzie Delta plain. It is also present in the broad valley floors within the western mountains (Mackenzie and Richardson mountain ranges), and in sandy deposits of the Pleistocene Mackenzie Delta of Richards Island, Yukon Territory coastal plain, and parts of Tuktoyaktuk Peninsula.

Low ice-content permafrost

Containing generally less than 5% visible ice, low ice-content permafrost is found primarily in three situations: 1) the foothills of the western mountains, south of 64°N, and including the eastern slopes of McConnell Range; 2) the mountainous terrain of the northern Canyon Ranges and the Richardson Mountains, north of 64°N; and 3) areas of coarse-grained and well drained deposits within the Interior Plains; ridged moraine in the southeast of the study area; rock outcrop in hills; coarse alluvial deposits in river valleys in the central part of the study area; and glaciofluvial deposits in the Tuktoyaktuk coastlands.

Negligible ice-content permafrost

Permafrost which contains no visible ice is described herein as having negligible ice content. This category is present primarily in the southern parts of the Canyon Ranges and Franklin Mountains.

Variable ice-content permafrost

There are also extensive areas of permafrost with widely variable ice contents. These comprise 1) subregions where distinct patches of high, moderate, low, and even negligible ice-content permafrost are intermixed, and 2) areas of primarily hummocky and ridged moraine with low ice-content ground on the tops of hummocks and ridges, and high ice-content ground in the intervening depressions. The former category is widespread in the southern part of the study area, particularly in areas of alluvial and lacustrine deposits associated with glacial Lake McConnell. These areas are found alongside the valley of Mackenzie and Liard rivers, to the south of 64°N. The second category occurs in patches on the plains. It is most common south of 64°N, but smaller areas are found north to the Arctic coast.

Areas free of permafrost

There are also a number of areas that are essentially free of permafrost and so contain no perennial ground ice. Most occur in the area either side of the Mackenzie River, between Mills Lake and the confluence of Liard and Mackenzie rivers.

ICE WEDGES, MASSIVE ICE, AND PINGOS

Large bodies of ice, occurring within ice-bonded permafrost and having major dimensions ranging from a few metres to many hundreds of metres, occur in three main forms: as ice wedges, as massive icy beds, and as the cores of pingos.

Ice wedges

These occur widely throughout the northern part of the area, but only sporadically in the southern part. Active ice wedges (downward tapering bodies of ice which are growing where meltwater fills thermal contraction cracks) are probably confined to the northernmost part of the region, generally north of 68°N. Even in this area, many ice wedges do not crack each year. In the northern part of the study region, the general distribution of active ice wedges can be gauged from the occurrence of intersecting networks of frost polygons or ice-wedge polygons. The distribution of inactive ice wedges is poorly known, however, as most have no surface expression.

a

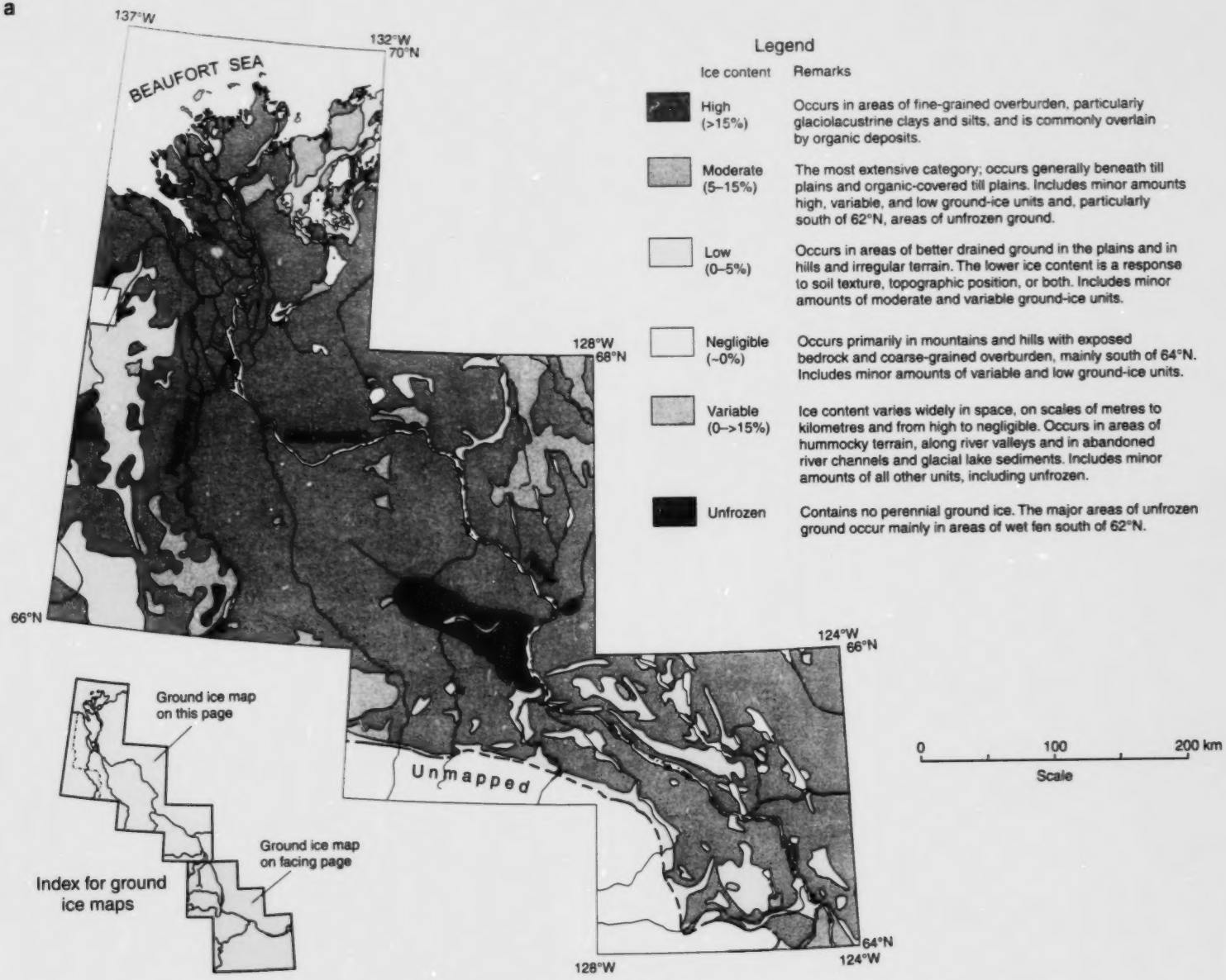
137°W
132°W
70°N
BEAUFORT SEA
128°W
68°N
66°N
124°W
66°N
Unmapped
64°N
128°W
124°W
Ground ice map on this page
Ground ice map on facing page
Index for ground ice maps
Legend
Ice content
Remarks
High (>15%)
Occurs in areas of fine-grained overburden, particularly glaciolacustrine clays and silts, and is commonly overlain by organic deposits.
Moderate (5–15%)
The most extensive category; occurs generally beneath till plains and organic-covered till plains. Includes minor amounts high, variable, and low ground-ice units and, particularly south of 62°N, areas of unfrozen ground.
Low (0–5%)
Occurs in areas of better drained ground in the plains and in hills and irregular terrain. The lower ice content is a response to soil texture, topographic position, or both. Includes minor amounts of moderate and variable ground-ice units.
Negligible (~0%)
Occurs primarily in mountains and hills with exposed bedrock and coarse-grained overburden, mainly south of 64°N. Includes minor amounts of variable and low ground-ice units.
Variable (0–>15%)
Ice content varies widely in space, on scales of metres to kilometres and from high to negligible. Occurs in areas of hummocky terrain, along river valleys and in abandoned river channels and glacial lake sediments. Includes minor amounts of all other units, including unfrozen.
Unfrozen
Contains no perennial ground ice. The major areas of unfrozen ground occur mainly in areas of wet fen south of 62°N.
0
100
200 km
Scale

Figure 1. *Maps showing* ***a****) distribution of ground ice, and* ***b****) extent of permafrost. Index map shows relationship between the north on south parts of the ground-ice map. The permafrost map covers the same area as the ground-ice map.*

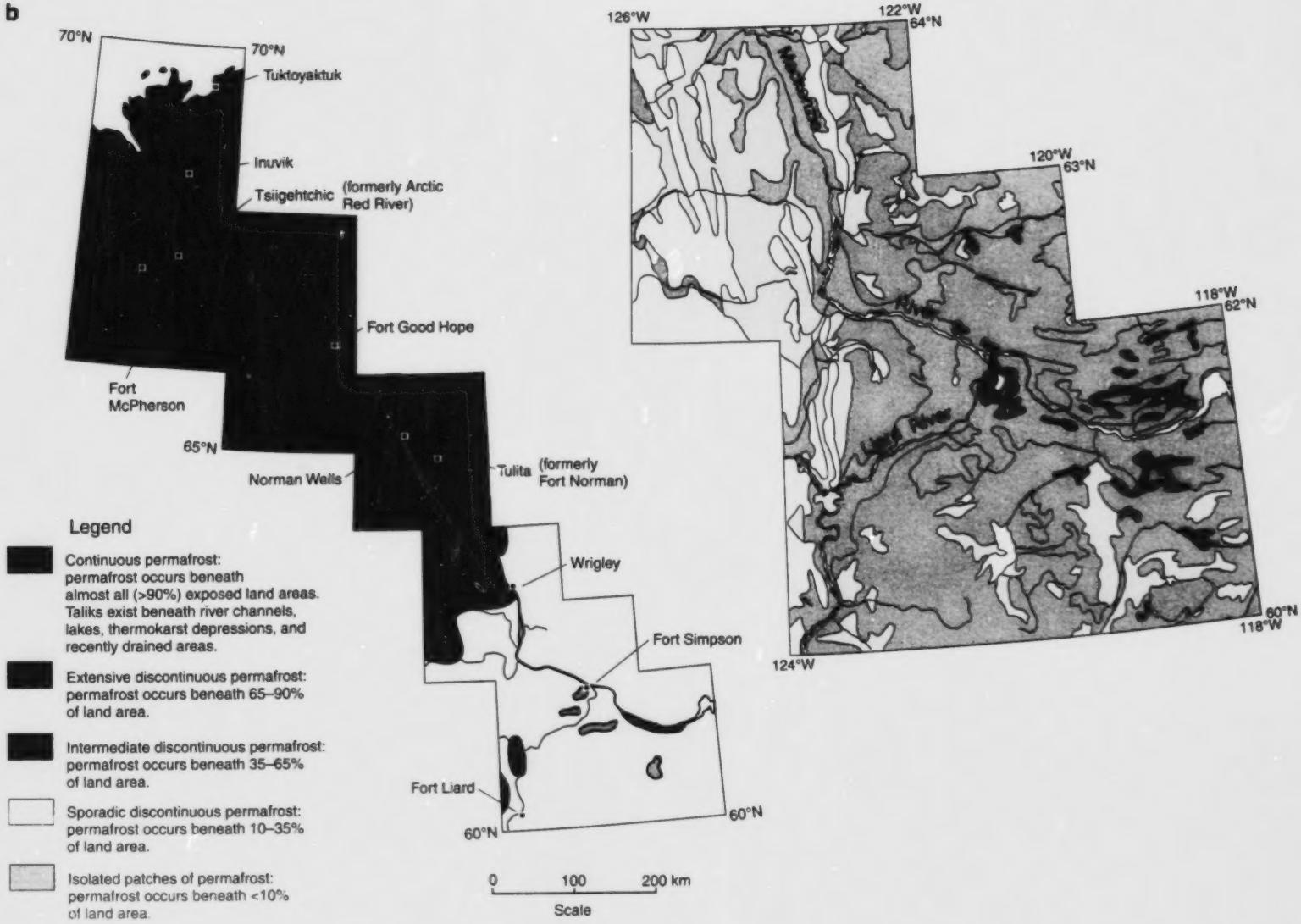

***Figure 1** (cont.)*

Massive icy beds

Massive beds of ground ice are found exclusively in the northernmost part of the region. They are well exposed in coastal sections of the Yukon Territory coastal plain, the islands off the Mackenzie Delta, and the Tuktoyaktuk Peninsula, including the Eskimo Lakes. Their origin and distribution have been discussed by Mackay (1963, 1971, 1973), Rampton and Mackay (1971), and Mackay and Dallimore (1992). It is believed that much of the higher ground of the Pleistocene Mackenzie Delta, particularly Richards Island and the Tuktoyaktuk Peninsula, owes its elevation to thick, extensive, tabular sheets of ground ice (Rampton and Mackay, 1971; Rampton and Walcott, 1974). Mackay (1963, Fig. 21) included a map of 'ground ice slumps' (now also referred to as 'retrogressive thaw slumps', or 'bimodal flows'); as many of these exposed massive ground ice, Mackay's map gives some indication of the wide extent of massive ice deposits in the area east of the Mackenzie Delta. It is possible that some of these ice masses are relict glacial ice; others may be buried deposits of other surface ice, such as river icings, snowbanks, or lake ice. Mackay's map of ground-ice slumps is reproduced as Figure 2.

Pingo cores

Pingos, large conical to hemispherical mounds or small hills with icy cores, also occur exclusively in the northernmost part of the region. Their distribution was also mapped by Mackay (1962, Fig. 1). Mackay's map of pingo distribution in the Tuktoyaktuk coastlands is reproduced as Figure 3. The Tuktoyaktuk Peninsula and adjacent areas include the world's largest concentration of pingos — over 1450 in an area 120 km long by 50 km wide. Their distribution and origin have been reported on extensively by Mackay (1962, 1963, 1979).

GEOTECHNICAL CONSIDERATIONS

Throughout the region, ground ice is generally within a few degrees of its melting point, often within a fraction of a degree of 0°C, and is susceptible to surface disturbance and slope failure. On level ground, in fine-grained, high ice-content deposits, surface disturbance commonly leads to the development of thermokarst, extensive subsidence of the ground surface and the growth of thaw lakes. On slopes, deeper thawing leads to slope failure, initially in the form of skin flows or

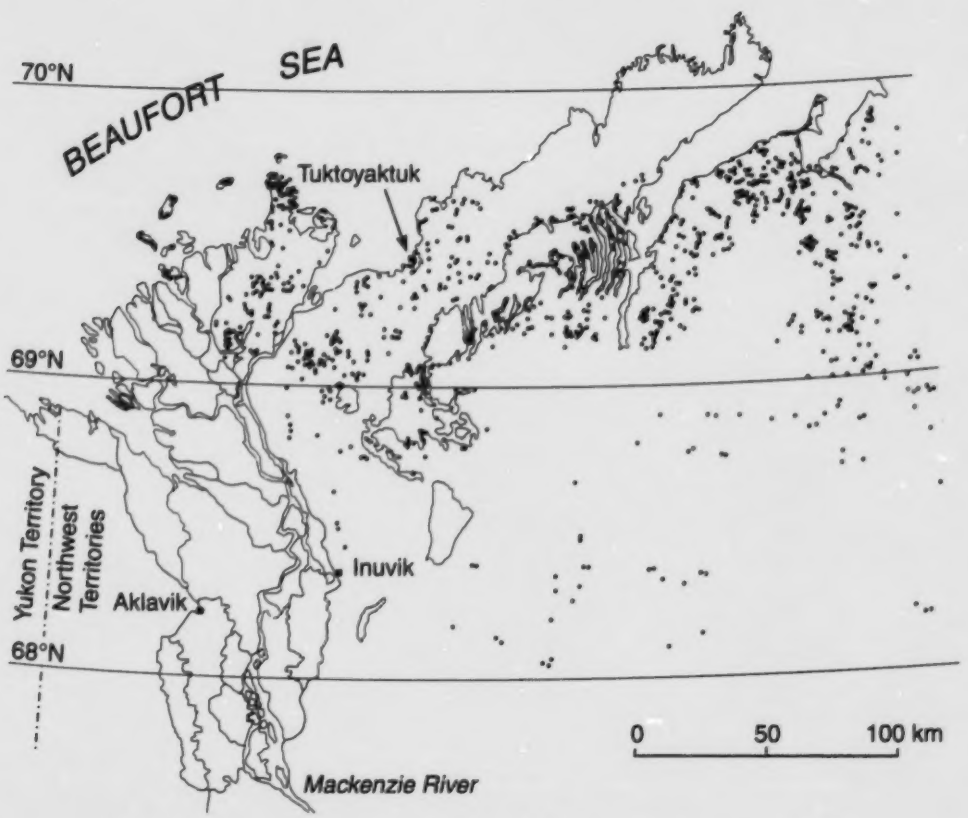

Figure 2. *Location where massive ground ice has been exposed in slumps. Where dots are clustered, they represent one or more slumps. Where dots are dispersed, they usually represent only one slump* (after *Mackay, 1963).*

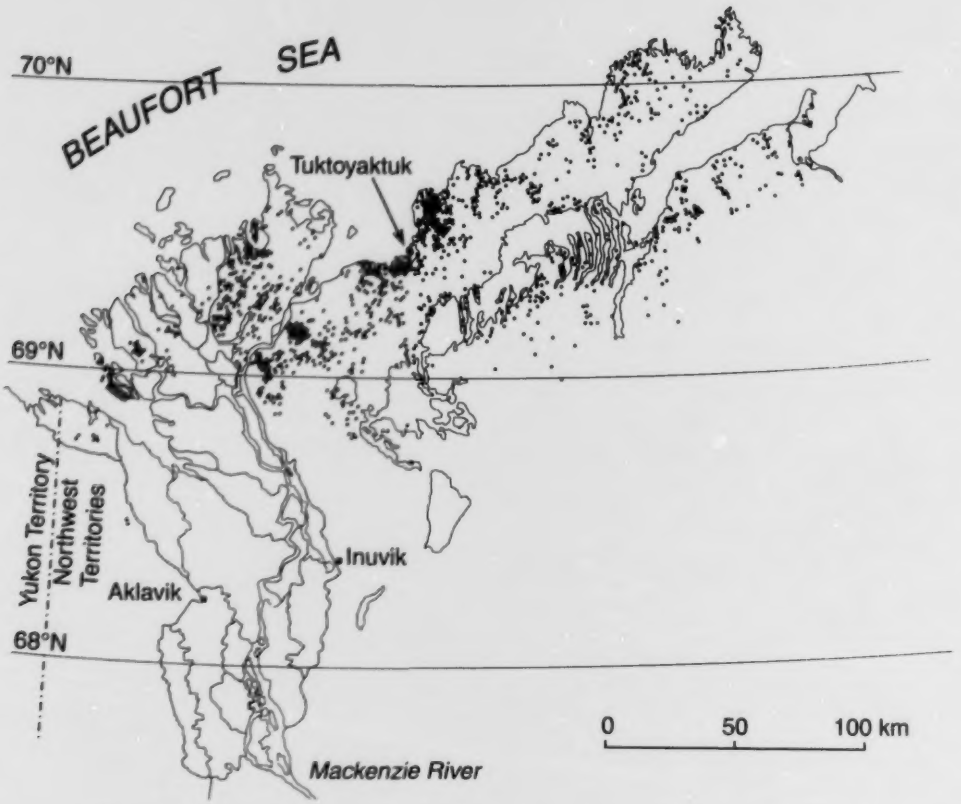

Figure 3. *Distribution of pingos as mapped from airphotos* (after *Mackay, 1963*).

active-layer failures. If such shallow slope failures expose massive ground ice, retrogressive thaw slumps can develop. Along the Beaufort Sea coast are sections of cliff coastline cut in frozen, unconsolidated sediments. Such materials are susceptible to thermal-niche development and failure as thermoerosional falls. Thermal niches can develop very rapidly under storm conditions in the late summer.

Saline permafrost is known to exist at the Tuktoyaktuk, Inuvik, Tsiigehtchic, and Fort McPherson settlements, and also at sites on Richards Island (Hivon et al., 1989; Hivon, 1991), all in the northernmost parts of the region. It may be assumed that saline permafrost underlies other parts of the area. The origin of this salinity has not been determined conclusively. In part it may result from the dissolution of saline rocks; elsewhere, particularly in the northern part of the region, it may result from the incorporation of seawater into the permafrost body. The salinity of the ground is important because high salinities have negative effects on the mechanical properties (strength, creep, etc.) of soils, and so may necessitate modifications to engineering designs.

IMPLICATIONS OF CLIMATE CHANGE

Under conditions of global climatic change, permafrost and ground-ice conditions will change. Climate warming will lead to an overall reduction in the extent of permafrost and a melting of ground ice, with a general northward shift in the southern limits of the permafrost zones. A cooling climate will cause an increase in the extent of permafrost and, with time, an increase in the amount of ground ice, particularly structure-forming ice. As discussed in Dyke and Brooks (2000), there is evidence that the climate of the Mackenzie valley has warmed in recent decades. Currently, global circulation models (GCMs) predict a general warming of climate worldwide by several degrees Celsius. Most GCMs also predict that this warming will occur first and to a greater than average degree in arctic and subarctic regions — the very areas encompassed by the Mackenzie valley. Thus, climatic warming of the past century in this area is anticipated to continue.

The implications of a warming scenario will therefore be discussed in more detail than those associated with climate cooling. It should be noted, however, that even under a

general warming of the climate, there will be localities where the ground-surface temperature becomes cooler and permafrost therefore cools and thickens.

A general warming of the climate will have a significant effect on the extent and distribution of permafrost and ground ice, and on permafrost-related geomorphic processes within this region. In general, as the ground warms, the active layer will increase in depth and the icy permafrost immediately below the active layer will thaw. On level to gently sloping sites, thermokarst pits and ponds will develop in areas of poor soil drainage. On slopes, the incidence of shallow slope failures will increase in extent and frequency. Where such active-layer failures (also referred to as skin flows) expose massive ground ice, retrogressive thaw slumps will develop. Ice wedges pose a special difficulty in that inactive wedges often have no surface expression; this can lead to problems for the construction and maintenance of roads, airstrips, and pipelines. As the number of active wedges decreases and more of the inactive ones begin to thaw, these problems are likely to increase significantly. In areas near coasts and rivers, deepening of the active layer and thaw of ice wedges will lead to the drainage of small lakes and ponds. As long as the local air climate is conducive to the continued existence of permafrost, permafrost will begin to develop in these newly exposed areas of land.

Climate warming will gradually lead to changes in vegetation and probably in the incidence of wildfire. Both of these changes will have effects on permafrost conditions, in the long term, where vegetation changes lead to further changes in the relationship between air climate and soil climate, and in the short term, when wildfire further accelerates the incidence of slope failure and the rate of vegetation change locally.

On a site-by-site basis, the effects of climate warming can be assessed qualitatively with a fair degree of accuracy. The general low thermal conductivity and diffusivity of earth materials, coupled with the high latent heat required to melt ice, lead to a lag in time between warming of the air and ground surface and effects on permafrost actually developing. For deep permafrost, this lag time can amount to many thousands of years; deep relict permafrost, which developed during the Pleistocene glacial periods, still exists beneath the Beaufort Sea continental shelf, immediately to the north of the study area. All this shows that while the effects of climate warming on permafrost conditions may be predictable, the timing is quite uncertain; it is amenable to modelling, but only on either a detailed site-by-site basis or in a very general way.

These changes have clear implications for the infrastructure, settlements, and economic activities of the population of the region. Increases in the incidence of slope failures, particularly along river banks, are likely to increase the silt load in rivers, and so affect domestic and commercial fisheries. A deepening of the active layer may negatively affect the stability of roads, airstrips, pipelines, and the foundations of buildings and other structures. Of particular concern will be the stability of tailings-pond dams, water reservoirs, oil tanks, etc., where catastrophic structural failure could lead to other, serious environmental problems.

Finally, there is concern that thawing of permafrost could result in the release to the atmosphere of large quantities of carbon dioxide and methane, which would accelerate the greenhouse effect and increase the rate of future climate warming. These gases would come from the decay of organic material, currently preserved by being frozen, and from the release of gas hydrate presently trapped beneath permafrost.

Any general cooling of the climate would be expected to reverse most of the effects already discussed. Permafrost would increase in extent and, eventually, in thickness. The active layer would be thinner and high moisture-content soils at its base would be incorporated into the permafrost, increasing the iciness of the uppermost part of the permafrost. Slopes would be come generally more stable; ice wedges would be rejuvenated; and thermokarst development would be retarded. Most of these effects would have only limited implications for settlements or infrastructure.

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Surficial geology, subsurface materials, and thaw sensitivity of sediments

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Abstract: The surficial geology of the Mackenzie valley and adjacent areas is presented in 1:1 000 000 scale compilation maps and is accompanied by geotechnical data from 56 boreholes chosen to represent typical stratigraphy of each major geological unit at several locations along the valley. For each of the representative boreholes, the potential for settlement of the ground surface as frozen ground thaws is determined, based on the thaw strain calculated for each soil layer of the borehole. Those geological materials most sensitive to thaw settlement are lacustrine and morainal, ice-rich, fine-grained sediments, and also the ice-rich and potentially highly compressible peat bogs. Much of the Mackenzie region is underlain by these materials; thus extensive terrain is susceptible to the impact of warming climatic conditions.

Résumé : La géologie des dépôts superficiels de la vallée du Mackenzie et des régions adjacentes est représentée sur des cartes de compilation dressées à l'échelle de 1/1 000 000. À ces cartes, s'ajoutent des données géotechniques provenant de 56 sondages choisis pour établir la stratigraphie typique de chacune des principales unités géologiques en différents endroits de la vallée. On a déterminé pour chaque sondage représentatif le tassement possible de la surface du sol suite à un dégel, en se basant sur la déformation causée par le dégel de chacune des couches de sol traversées par le sondage. Les matériaux les plus susceptibles de se tasser suite à un dégel sont les sédiments lacustres et morainiques à grain fin et riches en glace, ainsi que les tourbières oligotrophes à contenu élevé en glace qui affichent un potentiel de compressibilité élevé. Ces matériaux composent la grande partie de la région du Mackenzie; c'est pourquoi cette vaste contrée risque de subir les effets d'un réchauffement climatique.

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INTRODUCTION

A thorough understanding of surficial sediments and their ice content in the Mackenzie valley is key to predicting the impact of climate change on geological processes such as ground subsidence and landsliding. In this paper, the surficial geology of the Mackenzie valley and adjacent areas is summarized, and geotechnical data are presented from selected boreholes chosen to represent typical stratigraphy of each major geological unit at several locations along the valley. For each of the representative boreholes, the potential for settlement of the ground surface as frozen ground thaws is calculated. Finally, general conclusions as to the thaw-settlement potential of the major surficial geology units are made.

SURFICIAL GEOLOGY

Methodology

Maps at 1:1 000 000 scale (Fig. 1a, b, in pocket) of the surficial geology of the Mackenzie valley, adjacent mountains, Mackenzie Delta, and adjacent Beaufort Sea coast were compiled from existing published and unpublished surficial geology maps. Of these, detailed geology at 1:250 000 scale of the northernmost area (68–70°N) was mapped by Rampton (1981a, b). The central area (64–68°N) was mapped at 1:250 000 scale by Duk-Rodkin and Hughes (1992a, b, c, d, e, f, g, h, 1993a, b, in press a, b). The geology of the area east of longitude 126°W and between 64°N and 66°N was compiled from various reconnaissance-level geology maps by Fulton (1970), Hughes (1970), Hanley (1973), and Hanley et al. (1975). For the southern area (60–64°N) the existing compilation map at 1:500 000 scale by Rutter et al. (1993) was augmented with peatland data derived from earlier 1:125 000 scale maps by Rutter and Boydell (1979, 1980a, b, c, 1981) and Rutter et al. (1980a, b, c, d, e, f)

At the 1:1 00 000 compilation scale, some details of the geology are lost due to the merging of geologically similar units and the disappearance of smaller units. Wherever possible, the maximum level of detail was retained. In a few areas, compilation was difficult because adjoining published maps were produced at different scales or different levels of detail. In these cases, some geological units end abruptly at the margin of the original map sheet (i.e. 64°N latitude or 126°W longitude); however, as the original sheets were mapped as part of large blocks using common legends, these boundary problems are limited.

A brief description of the origin, texture, thickness, morphology, drainage characteristics, permafrost, and ground-ice conditions of each map unit is provided in the legend accompanying the compilation map (Fig. 1a, b). This information is derived from legends accompanying the original surficial geology maps and the amount of detail, particularly with respect to permafrost and ground-ice conditions, varies accordingly. Thickness of permafrost is not included within permafrost conditions.

Surficial materials

Surficial materials are sediments deposited by ice, water, wind, and gravity during Quaternary time — the ice ages through to the present. The detailed glacial and deglacial history of the area is covered by Duk-Rodkin and Lemmen (2000). Permafrost currently underlies most of the Mackenzie valley; as a result, surficial deposits are generally frozen and frequently contain ground ice. A discussion of permafrost distribution and ground-ice types and conditions is presented by Heginbottom (2000). The general distribution and nature of surficial materials is summarized below.

Much of the Mackenzie valley has a thin and discontinuous to thick and continuous cover of glacial ground moraine (till) which, in places, blankets or veneers the underlying bedrock topography and in other areas imparts a hummocky, rolling, or streamlined morphology to the surface (Fig. 2). Till is a composite of clay-sized to boulder-sized materials; where the matrix is fine grained, ground ice is often present. Till underlies many of the other surficial units. Associated with the glacial sediments are more localized areas of glaciofluvial sand and gravel that were deposited by meltwater in, under, or near the glacier ice. These deposits either take the form of hummocky and ridged topography, or broad, commonly pitted, outwash plains and terraces. These coarser grained sediments, essential for construction, are usually well drained and less susceptible to ground-ice formation.

Thick deposits of glaciolacustrine and lacustrine silt and clay typify much of the central axis of the Mackenzie valley. They mark the position of temporary, large lake basins formed during deglaciation due to the combined effect of isostatic depression of the surface and the damming of drainage behind retreating ice margins. These fine-grained sediments are particularly susceptible to the formation of ground ice, and consequently thermokarst features are widespread. In places, the silt and clay deposits are covered by sand of raised deltas.

Following the drainage of the glacial lakes, the exposed lake beds were subjected to the action of wind and flowing water. In places, large dune fields, now stable, formed on the exposed lake bed. These dunes impede local drainage, creating conditions favourable to the accumulation of organic materials. Elsewhere, alluvial terraces were formed as the Mackenzie River and its tributaries eroded the lacustrine sediments in response to lower base levels. Alluvial sediments, locally ranging from silt to sand and gravel, cap many of these terraces. The Mackenzie River delivered sediments forming a delta on the Beaufort Sea — a process that continues to the present day. Modern alluvial deposition continues on the floodplains, and point bars and channel bars of the rivers.

Peatlands are thick accumulations of organic materials that overlie extensive areas of flat-lying, poorly drained lacustrine sediments. They are also found in depressions in ground moraine and, less frequently, other surficial geology units. In the Mackenzie valley, two classes of peatlands occur: fen, commonly unfrozen, and bog, generally frozen and ice rich (*see* Aylsworth and Kettles, 2000). Thermokarst is widespread in the latter.

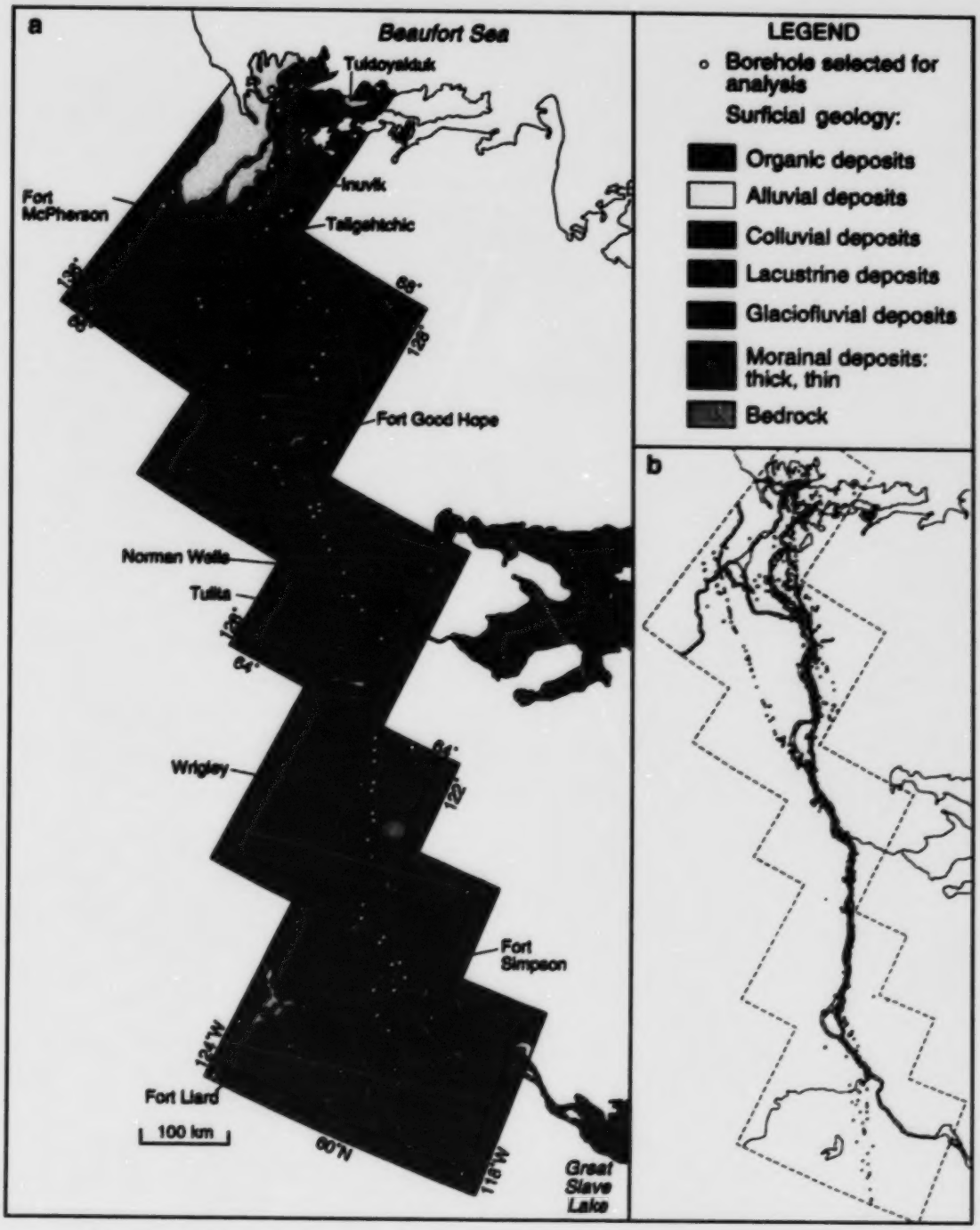

Figure 2. *a) Major surficial geology units in the Mackenzie valley and adjacent areas and location of the boreholes selected for analysis. For detailed geology see Figures 1a and 1b in pocket.* ***b)*** *Distribution of geotechnical boreholes in the Mackenzie valley.*

Colluvial sediments characterize the surface of much of the mountainous areas, although they also occur on steeper valley sides cut into more subdued landscapes. These materials result from the downslope movement of material under the influence of gravity at speeds ranging from a slow creep to rapid landslides. Colluvium is derived from weathered bedrock as well as the entire range of surficial deposits; the texture and ice conditions of colluvium reflect the original deposit. Fine-grained colluvial material may contain segregated ice. In some instances it is the thaw of ice-rich sediments which induces the downslope movement of material.

Fine-grained till and lacustrine sediments characterize much of the Beaufort coastal plain. Both contain massive bodies of ground ice and are especially prone to thaw-induced slope failures and thermokarst development. Eolian dunes are associated with several extensive areas of hummocky glaciofluvial sand and gravel. Although marine sediments are known to underlie the surface deposits in a few locations within the coastal plain, marine deposits are, for the most part, restricted to beaches and tidal ponds along the modern coastline. Alluvial deposits are present along the alluvial channels of the Mackenzie River system and on the Mackenzie Delta.

SUBSURFACE MATERIALS

Methodology

Selected boreholes were chosen from the 'Mackenzie Valley Geotechnical Database' to represent the typical stratigraphy of surficial geology units for several locations along the valley (Fig. 2a). The database (updated from GSC Open File 350 by Lawrence and Proudfoot (1976)) contains records of over 12 000 boreholes drilled for route selection and design of pipelines and roads, either proposed or constructed, in the Mackenzie valley since the 1970s. The boreholes thus are concentrated in corridors aligned with these routes (Fig. 2b).

The main criterion in selecting a representative borehole was an examination of the soil description. In addition, an effort was made to select boreholes that also contained sufficient geotechnical data to allow an estimate of the response to thaw (when the unit is described as generally frozen). Depths of selected boreholes are typically less than 15 m.

Borehole data

Borehole information is presented on Figures 1a and 1b. Note that some boreholes were situated in small material units that have been lost in map compilation. Thus, on the compilation maps, these boreholes now seem to lie within a different material unit (i.e. number 3, a borehole in glaciofluvial sediments, lies within a moraine plain unit). Figure 3 provides details on the nature and layout of the borehole data; the borehole identifier is that from the geotechnical database. Material properties are plotted with depth. Beginning with the first two columns at the left are a simplified lithology (*see* legend on Fig. 1a and 1b) based on the field description of the major soil layers, and the USC code (Unified Soil Classification system). The USC system (American Standard Test Method, 1990) is widely used by engineers and geologists for soil classification of samples collected from geotechnical boreholes. The USC codes are positioned at the approximate depth where samples were taken. When a sample is coded the same as an overlying one, an asterisk appears in the column at the sample depth. Note also that since the simplified lithology is based on a field description, and the USC description is generally based on laboratory analysis, there may be discrepancies between these two columns.

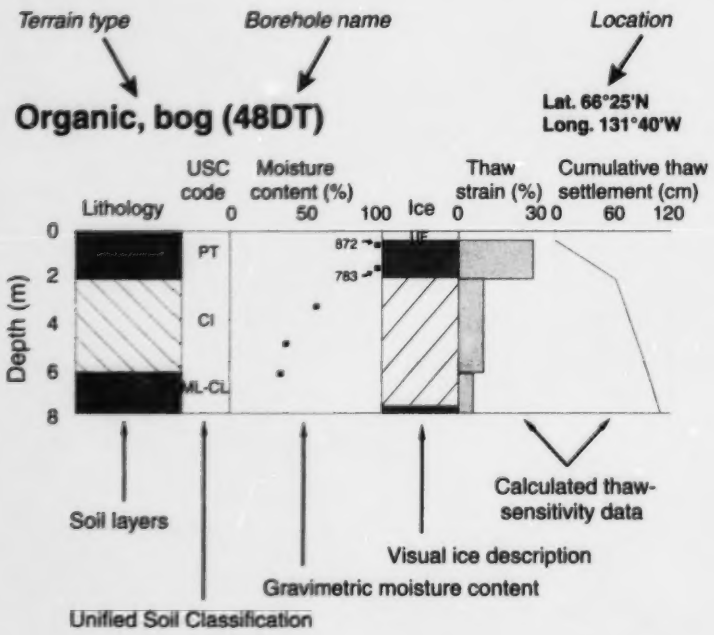

Figure 3.

Example of representative borehole log.

Moisture contents (expressed as per cent of dry weight) from borehole samples are plotted in the third column. Where the moisture content is greater than 100%, the value is indicated at the correct depth. The borehole logs of frozen sites generally include a description of the ground ice, based on the National Research Council classification system developed by Pihlainen and Johnston (1963). A record of ice encountered is contained in the fourth column. An explanation of colours and codes is provided in the legend. For a detailed statistical analysis of the occurrence of visible ground ice within the entire geotechnical database, *see* Heginbottom et al. (1978). For a quantitative examination of the percentage of visible ice along the Norman Wells pipeline, see Burgess and Lawrence (2000).

None of the borehole data should be presumed to represent conditions outside the borehole corridors. This is particularly true where there is a difference in altitude, because of differences in climate and associated changes in ground temperature. For example, at the southern end of the Mackenzie valley, along the Norman Wells pipeline, permafrost occurrence increases with elevation onto the Alberta plateau (*see* Burgess and Lawrence, 2000). On the other hand, in the vicinity of Norman Wells, permafrost may disappear with increase in altitude; permafrost exists at a site in the valley floor, but is not present at a site about 250 m higher in elevation (*see* Burgess and Lawrence, 2000). Such a difference near Norman Wells may be related to pronounced winter air-temperature inversions in the region. As well, broad regional trends in permafrost conditions related to general climate trends while local variability exists due to differences in site specific conditions such as aspect, snow cover, vegetation, thickness of insulating organic mat, and shifting stream channels (*see* Burgess and Smith, 2000; Dyke, 2000).

THAW CHARACTERISTICS OF SUBSURFACE MATERIALS

Methodology

An estimate of thaw settlement for each borehole was determined using the thaw strain calculated for each soil layer. The thaw strain is a measure of the per cent change in height or thickness of a soil unit upon thaw, and is a function not only of the volume change involved in melting the ice, but also of the compressibility of the soil layer. The compressibility is generally one to two orders of magnitude smaller than the volume change due to thaw, particularly for the relatively thin thickness (less than 15 m) of unconsolidated material under consideration here, and therefore it has not been included in the calculations. Thaw strains were calculated using methods reported in the literature for major soil groups (Speer et al., 1973; McRoberts et al., 1978; Interprovincial Pipe Line (NW) Ltd., 1982). These methods require a knowledge of sediment type, and of one of the following properties: total moisture content, ice content, or frozen bulk density. Moisture content is most likely to be available. Calculated thaw strains for each frozen soil layer are provided whenever possible in the fifth column of the borehole data plots. (Note: thaw strains are 0 at low moisture contents, because there is no excess ice present).

The calculated thaw strains were then used to determine the cumulative thaw settlement at each borehole resulting from thawing to the bottom of the hole (or to bottom of permafrost if the permafrost was thinner than the total borehole depth). Regardless of the source or rate of thermal disturbance leading to thaw, this cumulative thaw settlement provides an indication of how settlement at the surface would increase as thaw progressed. The cumulative thaw-settlement plots begin at the depth where the borehole is logged as frozen, or from the top of the layer with sufficient data to calculate thaw strain. Since many boreholes were drilled in winter, this depth is often at the ground surface and hence includes the active layer (i.e. the zone that freezes and thaws annually). A discussion of active layers in the Mackenzie valley can be found elsewhere in this volume (*see* Nixon, 2000).

The abbreviation 'NA' appears in the thaw-strain and thaw-settlement columns whenever these calculations are not applicable, i.e. if there is no permafrost at the site, if the moisture contents are too low, or if the data is insufficient to allow a calculation of these parameters. Note also that the depth-scale increments are identical for all boreholes, but that the horizontal scales for moisture content, thaw strain, and thaw settlement vary from site to site.

Thaw sensitivity of sediments

The thaw sensitivity data represent actual calculations for specific sites and not the 'average' value from all boreholes in the corresponding terrain unit in the area. Despite this, several general and noteworthy trends can be seen in the data presented. Thaw strains generally decrease with increasingly deep soil layers for the borehole depth ranges presented here. In the north, however, the reverse is frequently observed, often as a result of buried massive ice (*see* for example boreholes for lacustrine and glaciofluvial units in the 68–70°N area).

The calculated cumulative thaw settlement that would occur at each permafrost site under thaw depth increases to 3 m and 5 m are given in Table 1. The warming causing the thaw is not linked to any particular source, be it natural, climatic, or anthropogenic, nor to any particular duration. For a discussion of the rate and magnitude of thawing at select locations in response to different rates of climate warming in the Mackenzie valley, *see* Burgess et al. (2000).

Organic, moraine, and lacustrine units show the greatest thaw strains. There is also a tendency towards increasing thaw strain with increasing latitude, especially for lacustrine and moraine units. For example, 5 m of thaw in moraine in the 64–66°N zone could lead to 10–25 cm of settlement, but could result in 50–80 cm of settlement in moraine of the northernmost areas between 68–70°N. Frozen organic terrain on the other hand has high thaw strains wherever encountered. For example, 5 m of thaw within frozen bog could result in 50 to 120 cm of settlement. Alluvial terrain is frequently thawed, but when frozen, thaw sensitivity is greater in the 64–66°N zone due to the finer texture of sediments there. Glaciofluvial terrains, commonly coarse-grained sands and gravel, show low thaw strains and thaw settlements of

Table 1. Calculated thaw settlement (in centimetres) for the representative boreholes at different zones of latitude, using a thaw depth of 3 m and a thaw depth of 5 m.

Terrain unit	60–62°N		62–64°N		64–66°N		66–68°N		68–70°N	
	3 m	5 m	3 m	5 m	3 m	5 m	3 m	5 m	3 m	5 m
Organic (fen)	uf	uf					***38***	***38***		
Organic (bog)	48	50	86	117	77 73	90 120	48 70	49++ 87		
Alluvial fan			uf	uf			17	18		
Alluvial plain			7	7+	55	80	uf	uf	20	38
Alluvial terrace	uf	uf	10	***16***	31	43			uf	uf
Colluvial complex			15	20++	uf	uf			9	25
Eolian ridges	uf	uf	uf	uf			13	22		
Lacustrine plain	uf	uf			19	21	49	66		
Lacustrine plain (fine-grained sediment)			8	17	23	26			20 33	150+ 50
Lacustrine plain (coarse-grained sediment)			uf	uf						
Lacustrine complex					52	64				
Glaciofluvial plain	uf	uf			2	7	2	8		
Glaciofluvial terrace			6	10++	10	25			6	12
Glaciofluvial, hummocky							0	0		
Glaciofluvial complex							0	5		
Moraine plain	uf	uf	18	20++	21	25	54	70		
Moraine, rolling							18	31	75 43	82 49
Moraine, drumlinoid plain			25	28++	8	11	2	3		
Moraine, hummocky			15	27	27+	I.D.	54	61		
Moraine blanket									38	52

Notes: Blank entries indicate this is not a major unit within the map zone or no borehole fell within this unit.
Values in ***bold italic*** indicate that this is a maximum value for the borehole, since no permafrost remains below that depth.
'+' indicates that the borehole is less than 5 m deep and therefore the settlement reported is a minimum value.
'++' indicates that the value has been extrapolated from the data, e.g. where the hole is slightly shallower than 3 or 5 m.
'uf' indicates the terrain unit is a major one on the map sheet, and the borehole is unfrozen.
'I.D.' indicates insufficient data to estimate, e.g. hole too shallow.

less than 25 cm for a 5 m thaw depth. However, should massive ice underlie glaciofluvial and lacustrine units, greater thaw strains and settlement would be anticipated.

SUMMARY

The response of terrain to changing climatic conditions has particular significance for the design, construction, and maintenance of infrastructure and buildings on permafrost soils in the Mackenzie valley. Thick deposits of glaciolacustrine and lacustrine silts and clays, fine-grained tills, and thick accumulations of organic materials (especially in the south), are characteristic of much of the surficial materials of the Mackenzie valley. These subsurface materials are generally frozen and frequently ice rich. The presence of this excess ice causes these soils to be sensitive to climate warming — when subjected to thaw they will undergo some amount of thaw settlement for a period of time. The subsurface data presented in this paper quantify the potential thaw settlement of the major terrain units, while the accompanying figures show the distribution of the most thaw-sensitive materials. Those regions most sensitive to thaw settlement will be extensive areas of lacustrine and morainal, ice-rich, fine-grained sediments, and the ice-rich and potentially highly compressible peat bogs.

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Distribution of peatlands

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Abstract: The Mackenzie valley is characterized by vast peatlands dominated by two classes, bog and fen; the former is generally frozen and ground ice is common, the latter is unfrozen. Large peatlands are associated with broad featureless glaciolacustrine or till plains while smaller ones occur in low-lying areas between till or eolian ridges. Based on the compilation map, approximately 27 600 km^2 of bog and 10 000 km^2 of fen are estimated to occur in the Mackenzie valley. Although fens comprise only about 27% of the peatlands of the region as a whole, the proportion of fen to bog increases greatly in some areas. For example, close to Fort Simpson the ratio of fen to bog is 63:37. Peat thickness, based on information from the 'Mackenzie Geotechnical Borehole Database', increases towards the south.

Résumé : La vallée du Mackenzie est caractérisée par la présence de vastes tourbières qui sont attribuées principalement à deux classes : les tourbières oligotrophes (bogs) et les tourbières minérotrophes (fens). Les tourbières oligotrophes sont généralement gelées et contiennent de manière habituelle de la glace de sol, alors que les tourbières minérotrophes ne sont pas gelées. Les grandes tourbières sont associées à de vastes plaines de sédiments glaciolacustres ou de till dépourvues d'éléments géomorphologiques, tandis que les petites tourbières occupent des terres basses entre des crêtes de till ou de sédiments éoliens. Selon la carte de compilation, il existe environ 27 600 km^2 de tourbières oligotrophes et 10 000 km^2 de tourbières minérotrophes dans la vallée du Mackenzie. Les tourbières minérotrophes ne constituent que 27 % environ des tourbières de la région; cependant, leur proportion augmente de façon marquée dans certains secteurs. Par exemple, près de Fort Simpson, le rapport entre les tourbières minérotrophes et les tourbières oligotrophes est de 63/37. L'épaisseur de la tourbe, d'après les données géotechniques recueillies dans les sondages, augmente vers le sud.

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INTRODUCTION

Peatlands are areas where the soils consist of partially decomposed organic materials dominated by mosses and sedges and, to a lesser degree, shrubs and trees. The Mackenzie valley is characterized by vast peatlands across both the continuous and discontinuous permafrost zones. The presence or absence of permafrost in these peatlands has important implications for predicting ground behaviour in the event of continued climate warming. Thaw subsidence of the surface and release of greenhouse gases are two possible consequences of the thawing of permanently frozen peatlands. A knowledge of the distribution and nature of the peatlands is needed before potential effects of warming can be assessed. Accurate mapping of peatlands and knowledge of peat thicknesses, ground-ice content, and ground temperatures are necessary for predicting future changes in peatlands. The purpose of this paper is to show the distribution of peatland types and to discuss the thickness of peat and the permafrost conditions in the peatlands of the Mackenzie valley.

THE NATURE OF PEATLAND IN THE MACKENZIE VALLEY

Two major classes of peatland, bog and fen, are found in the Mackenzie valley (Tarnocai et al., 1995). Each has distinctive vegetation assemblages, morphologies, water regimes, and thermal conditions, with bog being typically frozen and fen unfrozen (Table 1).

Peatlands are characterized by layers of poorly to moderately decomposed organic materials at the surface and more thoroughly decomposed materials near the base. Macrofossil analysis of peat shows that bogs commonly developed from fens, which in turn, formed as organic materials accumulated in lakes or poorly drained depressions. In many peat plateaus only the top 50 cm is *Sphagnum* peat, covering fen peat (Zoltai and Pollett, 1983). Finally, collapse scars in degrading peat plateaus may develop into fens.

Bogs and fens take a number of distinct forms. Within the Mackenzie valley, the most prevalent bog form is the peat plateau, which is elevated 1–2 m above adjacent fens and covers a few square kilometres to several hundred square kilometres (Fig. 1, 2; Zoltai and Pollett, 1983). Another form is polygonal peat plateau, which is similar to a peat plateau, but contains polygonal ice-wedge trenches, the product of winter frost cracking. In the southern part of the Mackenzie valley, some areas of unfrozen bogs, formed from degraded peat plateaus, have been noted (S.D. Robinson, pers. comm., 1997). Also present are flat bogs and palsas; the first are featureless surfaces occurring in broad, poorly defined depressions, and the second are frozen mounds of peat, less than 100 m in diameter, which rise 1–7 m above the surrounding fen.

The common fen forms are the horizontal fen and northern ribbed fen. The former is characterized by a very gently sloping, featureless surface, which occupies broad, often poorly defined depressions. The latter have low, narrow peat ridges that cut across the fen at right angles to the direction of water movement, and form gentle arcs across the fens

Table 1. Comparison of physical and chemical characteristics of bogs and fens in the Mackenzie Valley, based on legend notes accompanying regional surficial geology maps.

FEATURE	BOG	FEN
Topography and morphology	Flat or very gently inclined plains; usually raised about 1 m above surrounding fens, or as peat filling slight depressions in the landscape. Commonly pocked with numerous steep-sided depressions (collapse scars) occupied by thaw ponds or small fens, or by steep-sided trenches that act as seepage channels connecting the depressions	Flat to very gently inclined plains. Commonly featureless plain; some have reticulate network of low (<50 cm) ridges
Typical peat thicknesses	64°-68° = 2–4 m thick; 60°-64° = 1.5–7 m thick	64°-68° = 2–3 m thick 60°-64° = 2–3 m thick
Surface vegetation	Cover of sphagnum moss and heath shrubs. May be open or forested by scattered to closed cover of black spruce.	Cover dominated by sedges; grasses and reeds may be found in local pools. May be open or forested by shrubs and scattered tamarack or black spruce trees.
Drainage pattern	Poor; numerous small ponds	Drainage mainly by subsurface seepage. Water table at surface during summer months; occasional small pools.
Acidity	Surface peats and waters are strongly acidic.	Fen peats and waters are less acid than in bogs
Permafrost	Commonly frozen 0.3-0.5 m below the surface; peat in wet depressions may be unfrozen to depths of more than 1 m. Segregated ice widespread within the peat (up to 80%) and in underlying sediment. Growth of segregated ice elevates surface of bog over surrounding fen; ice degradation forms collapse scars.	Generally unfrozen. In the south some relic permafrost may occur at depths exceeding 3 m or more. In the north, channel fens and small fens within bog complexes may be underlain by permafrost at depths exceeding 0.5 m.

(Fig. 2). The ridges are better drained and may be underlain by permafrost. There are several other forms, defined by their topographic location (basin, stream, channel, and shore fens) or permafrost history (collapse scar fens; *see* Fig. 1; Zoltai et al., 1988a, b).

Canada has been divided into wetland regions where specific wetland forms develop in locations that have similar topography, hydrology, and nutrient sources (Zoltai, 1988). Three wetland regions, High Boreal, Low Subarctic, and High Subarctic (Fig. 3), are contained in the Mackenzie valley. High Boreal wetland is characterized by small wooded peat plateaus, flat bogs, and palsas, with fens, including northern ribbed, collapse scar, basin, and shore forms (Zoltai et al., 1988b), whereas Low Subarctic wetland is characterized by peat plateaus, palsas, and northern ribbed, horizontal, channel, and shore fens (Zoltai et al., 1988a). Polygonal peat plateaus with local basin fens and shore fens are the principal forms of High Subarctic wetland (Zoltai and Pollett, 1983). In the High Boreal and Low Subarctic wetland regions, permafrost is associated with peat plateaus and palsas, but fens and their underlying sediments are unfrozen. In the High Subarctic region, permafrost underlies all peatlands, except portions of shore fens located within a few metres of a lake (Zoltai and Pollett, 1983). In many places in all three wetland regions, fens, and bogs are closely associated.

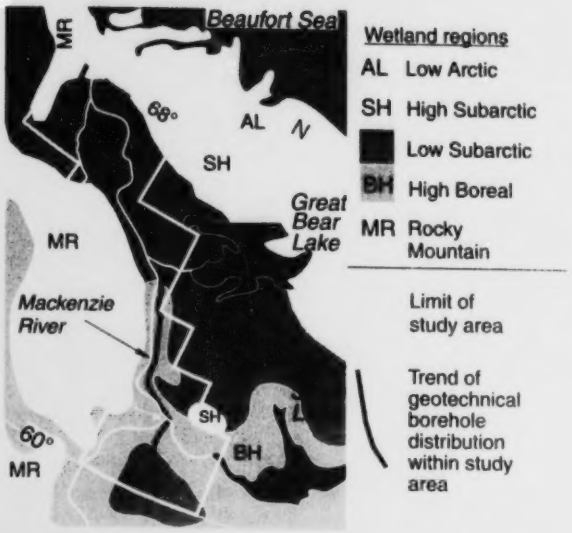

Figure 3. *Wetland regions* (from *National Wetland Working Group, 1988) and location of geotechnical boreholes within study area.*

Figure 1. *An isolated, round peat plateau, 225 m in diameter, within a fen (62°43'N, 121°06'W). A collapse-scar fen, approximately 30 m wide and 60 m long has formed within the peat plateau. Photograph by S.D. Robinson. GSC 2000-026A*

Figure 2. *Peat plateau, marked with collapse scars (right), and northern ribbed fen (left) approximately 100 km north of Fort Simpson. Photograph by S.D. Robinson. GSC 2000-026B*

DISTRIBUTION OF PEATLANDS IN THE MACKENZIE VALLEY

Peatlands, delimited mainly by airphoto interpretation, are shown as distinct terrain units on surficial geology maps of the Geological Survey of Canada. Due to the prevalence of peatlands in the Mackenzie valley, surficial mappers have depicted not only the boundaries of major peatlands, but have also indicated on the maps the extent of more discontinuous, minor peatlands. On the published maps, discontinuous peatlands are shown either as minor components of other surficial units (e.g. unit 'siLp-pO', meaning silty lacustrine plain with 16–49% of the surface covered with peat bog in Rutter et al. (1980c)) or as symbol overlays covering parts of other surficial units (e.g. Duk-Rodkin and Hughes, 1992f). Where peat exists solely as a veneer, not masking the underlying sediments, it is not mapped. The accompanying map (Fig. 4, in pocket), which shows the distribution of peatland in the Mackenzie valley, was compiled and generalized from 1:125 000 and 1:250 000 scale, regional surficial geology maps (Fulton, 1970; Hanley, 1973; Hanley et al., 1975; Rutter and Boydell, 1979, 1980 a, b, c, 1981; Rutter et al., 1980a, b, c, d, e, f; Duk-Rodkin and Hughes, 1992a, b, c, d, e, f, g, h, 1993a, b, in press a, b).

It is estimated that there are approximately 27 600 km^2 of bog and 10 000 km^2 of fen in the Mackenzie valley, based on Figure 4. Because the map is a generalization, the land surface represented by individual map-unit boundaries is unlikely to be covered completely by peatlands. The authors estimate, for the purposes of this calculation, that 1) 70% of the surface area of the 'extensive peatland' units and 20% of the 'discontinuous peatland' units are actual peatlands, and 2) the ratio between the two peatland classes is 2:1 for bog-dominant or fen-dominant mixed peatland units, and 1:1 for undifferentiated mixed peatland units. In general, large peatlands have formed on the extensive glacial-lake plains which cover the central axes of the Mackenzie valley and several tributary valleys. Where eolian deposits cover the glacial-lake sediments, peat has commonly accumulated in low-lying areas between sand dunes. Peatlands have also developed on the broad, featureless surfaces of till plains or in low-lying areas between drumlins or in depressions in hummocky moraines. Smaller accumulations of peat are found in depressions in alluvial or glaciofluvial deposits. In the mountainous areas bordering the plains and plateaus of the Mackenzie valley, peatlands are restricted to small accumulations on valley bottoms.

In the lower Mackenzie valley (64–68°N), which lies mainly within the Low Subarctic region, there are two major peatlands (Fig. 4). The first, dominated by bog, crosses the Arctic Red River and is adjoined by other large areas characterized by a discontinuous cover (<50%) of fen and bog. The second prominent peatland, dominated by bog or a mixture of bog and fen, is in the vicinity of the Great Bear River. Elsewhere in the lower Mackenzie valley, peatland distribution is more sporadic. Many long, narrow, stream fens occupy small valleys in the northeast, but they are too small and discontinuous to be shown on Figure 4.

In the upper Mackenzie valley (60–64°N), prominent peatlands dominated by fen lie in the central axis of the valley in the vicinity of Fort Simpson and eastwards (Fig. 4); these peatlands overlie glacial-lake plains and till plains within the High Boreal region. At higher elevations, away from the central axis of the valley, the till plains are characterized by extensive peatlands consisting of bog or of mixed peatland (bog with lesser amounts of fen). These bogs lie mainly within the Low Subarctic region.

Although fens constitute about 27% of the peatlands of the Mackenzie valley as a whole, the proportion of fen to bog increases greatly in some areas. For example, close to Fort Simpson the ratio of fen to bog is 63:37 (northwest quarter of NTS 95 H).

PEAT THICKNESS BASED ON BOREHOLE DATA

Typical peat thicknesses based on legend notes accompanying regional surficial geology maps are shown in Table 1. Other scientists have investigated peat depths associated with specific forms of bog and fen in the different wetland regions in the Mackenzie valley. In the Low Arctic region, peat is commonly 1.5 m thick under high-centre polygons, but only 50 cm under polygon fens (Tarnocai and Zoltai, 1988). Peat has accumulated in excess of 2 m, and up to 4 m thick, in peat-plateau bogs of the Subarctic region, while thicknesses of 0.4 m to an average of over 3 m have been recorded in channel fens, and 0.3–1.5 m in veneer bogs and collapse-scar fens (Zoltai et al., 1988a). In the Boreal region, peat thicknesses in peat-plateau bogs are commonly in excess of 2 m, but seldom more than 5 m, while accumulations in palsas may be over 4 m. In the same region, depths in excess of 1 m are reported in northern rib fens, 1 m to more than 6 m in basin fens, and 1–2.5 m in spring fens (Zoltai et al., 1988b).

The following section is not limited to mapped areas of peatland, but applies to the surface organic layer which may be present in a variety of terrain types. Most of the information is derived from the 'Mackenzie Geotechnical Borehole Database' (updated from Lawrence and Proudfoot, (1976)). This database contains records of over 12 000 boreholes drilled along the Mackenzie valley between 60°N and the Beaufort Sea primarily for pipeline- and road-route selection and design. The database is a compilation of information supplied by private industries from boreholes drilled, logged, and sampled for geotechnical purposes using a variety of drilling or coring methods. Drilling generally occurred in winter from winter roads. Stratigraphy was commonly logged in 0.5 foot increments. In most records, peat and highly organic soils (a mix of organic and mineral sediments dominated by the former) were grouped as one unit because they have much the same impact on construction. For the remainder of this section, peat and highly organic soils will be referred to as organics. Where the surface unit was logged only as organic soils without mentioning peat, the records were not used.

The data set is not a completely objective sampling of the regional organic depths because the boreholes are located within narrow linear corridors (the planned pipeline and highway routes) that were situated to avoid extensive peatland areas as much as possible. Also, the majority of boreholes are located in the northern half of the valley and often their specific location was intended to test for aggregate accumulations. Nevertheless, the borehole data set provides a useful indication of regional organic-layer thickness and allows a comparison between organic thickness and various environmental factors (latitude, underlying sediment type, and wetland region).

A shallow organic cover is common over much of the Mackenzie valley. For example, approximately half (6686) of the boreholes in the database noted surface organic thicknesses of at least 0.06 m. From the database, only organic thicknesses of at least 1.5 ft. (0.46 m), the closest measurement to the defined minimum thickness for peatland of 0.4 m (Tarnocai, 1980), were selected for the following analysis. It is recognized that the minimal thickness quoted by Tarnocai refers to peat alone, whereas the data set used here includes highly organic soils within the organic unit. Surface organics, ranging from 0.46 m to a maximum of 14.6 m thick, were recorded for 2708 holes, approximately 23% of the database. It is possible that some or many of the 2708 boreholes with organic-layer thicknesses greater than 0.46 m may not be located in recognized peatland.

To investigate the relationships between thickness of organics and latitude, thickness measurements for the 2708 boreholes were grouped by 0.25° increments of latitude. Maximum thicknesses of organic materials recorded for boreholes are found north of 66.5°N and between 64.5° and 65.25°N, however the greatest mean (4.8 m) and median (4.2 m) thicknesses lie south of 61.5°N (Fig. 5). The trend of increasing thickness of organic sediments to the south is confirmed by ditch-wall records for the Norman Wells to Zama pipeline (*see* Burgess and Lawrence, 2000).

Of the 2708 boreholes selected, 19% occur in the High Boreal wetland region (BH), 80% occur in the main Low Subarctic wetland region (SL-2), and 1% in the isolated area of Low Subarctic (SL-1) lying just north of the provincial border (Fig. 3). Only a small part of the study area lies in the High Subarctic zone and very few boreholes were located there. Each borehole record in the database notes the genetic class (organic, morainal, lacustrine and glaciolacustrine, colluvial, alluvial, fluvial and glaciofluvial, other) of the encompassing surficial geology unit, based on maps available at the time of drilling. Table 2 summarizes organic-cover statistics for each surficial unit within each wetland zone.

Thickness of the organic cover is generally greater in areas that were mapped as organic terrain, which in many locations overlie lacustrine deposits, and also in areas mapped as lacustrine (Table 2). The flat surfaces and poor permeability of lacustrine sediments impede drainage, which in turn fosters peatland development. Depths of organic materials are lower in colluvial areas, where the ground surface is unstable. Thicknesses associated with other surficial units are variable and probably reflect morphological controls. For example, accumulation of organics is common in low areas between drumlins and flutes in morainal terrain. In alluvial terrain, peat accumulation may be cyclic with sediment deposition; buried peat units were eliminated from this survey. When thicknesses of the organic layer for each genetic class are compared by wetland region, there is a general trend of increasing peat depths (mean and maximum values) from north to south through the Low Subarctic-2 to High Boreal to Low Subarctic-1 regions (Table 2).

Table 2. Surface peat, including highly organic soils, for surficial geology units within three wetland regions: Low Subarctic-1 (SL-1), 60.00–60.75°N; High Boreal (BH), 60.75–63.75°N; and Low Subarctic-2 (SL-2), 63.50–68.00°N. Data set limited to records in which this unit ≥0.46 m. Extracted from 'Mackenzie Geotechnical Borehole Database' (updated *from* Lawrence and Proudfoot (1976)).

			Thickness of peat (m)		
Geological unit	Wetland zone	Number of boreholes	Mean	Median	95th percentile
Alluvial	SL-2	83	0.97	0.61	2.13
	BH	82	0.61	0.61	1.22
Colluvial	SL-2	91	0.63	0.61	1.31
	BH	32	0.86	0.61	1.83
Fluvial and glaciofluvial	SL-2	140	0.80	0.61	1.83
	BH	24	1.06	0.61	1.83
Lacustrine and glaciolacustrine	SL-2	299	0.99	0.61	2.44
	BH	138	0.99	0.61	2.74
Moraine	SL-2	905	0.88	0.61	1.83
	BH	166	0.99	0.61	2.90
	SL-1	17	3.00	2.90	4.90
Organic	SL-2	423	0.97	0.61	2.74
	BH	29	1.82	1.52	4.27
Other	SL-2	238	1.21	0.61	3.00
	BH	41	0.91	0.61	1.22

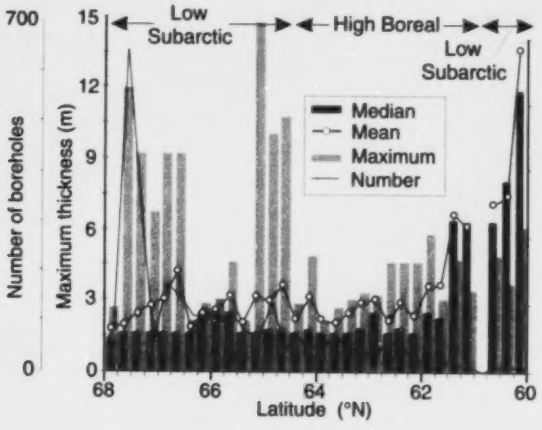

Figure 5. *Thickness of peat by 0.25° latitude grouping. Statistical summary from 2708 boreholes recording peat and highly organic soil cover ≥0.46 m (1.5 ft.) in the 'Mackenzie Geotechnical Borehole Database' (updated* from *Lawrence and Proudfoot (1976)). Note the three different y-axis scales.*

IMPLICATIONS OF CLIMATE CHANGE ON PEATLANDS

Thaw settlement

Because of the association of permafrost with bog, the presence of one peatland class as opposed to the other has important implications for predicting ground behaviour in the case of rising ground temperatures. Thaw subsidence resulting from degradation of segregated ice in bogs and the compressibility of thawed peat presents serious problems for construction and maintenance of highways, pipelines, buildings, and other infrastructure on the extensive organic terrain of the Mackenzie valley. Poor drainage in fens and the high compressibility and low strength of fen peat make fens unsuitable for any development. The greater frequency of frozen/unfrozen interfaces along any route in the southern part of the valley further complicates construction (*see* Burgess and Lawrence, 2000).

In Aylsworth et al. (2000), calculations of settlement in the event of thaw are presented for representative boreholes in various sediments. These data indicate that peat layers have moisture contents (up to 872% in ice-rich samples) and thaw strains (24–28%) that are greater than any mineral soil. This significantly affects the sensitivity of this type of ground to thawing. For example, with a thaw strain of 28%, a layer of peat 1 m thick would settle 28 cm when thawed. It should be

noted that ground ice is common in fine-grained sediments overlain by an insulating blanket of thick peat, and that significant thaw strains and thaw settlements have been calculated for these underlying deposits. Since ground ice is irregularly distributed in and under the peat of bogs, thaw settlement would occur nonuniformly. Thus thaw settlement, resulting from climatic or any other disturbance of the ground-temperature regime of peatlands, will have important consequences for development and maintenance of any infrastructure constructed upon the extensive organic terrain of the Mackenzie valley.

FUTURE EMISSION OF GREENHOUSE GASES

With climate warming, the release of greenhouse gases from peatlands will augment the degradation of permafrost. At present, the production of greenhouse gas is extremely limited within the frozen bogs, which constitute approximately 73% of the peatland area in the Mackenzie valley (*see* 'Distribution of Peatlands in the Mackenzie Valley', above). Within unfrozen fens, degrading peat may produce greater than 100 times more methane, a major greenhouse gas, than peat in nearby frozen areas, per unit area (Watson et al., 1990). For example, the mean seasonal flux of methane has been measured from some forms of peatland in the Fort Simpson area as follows: open poor fen, open fen, low shrub fen, and treed low shrub fen have a mean flux of 162.5, 63.5, 19.0, and 3.7 mg CH_4 per m^2 per day, respectively, and mean flux of a bog collapse scar is 109.5. In comparison, the mean flux of a frozen peat plateau is only -0.6 mg CH_4 per m^2 per day (Liblik et al., 1997). Emissions have also been measured for northern Manitoba, another boreal region in the discontinuous permafrost zone. In Manitoba, seasonal average fluxes from fens and collapse scars are 92–380 mg CH_4 per m^2 per day, whereas from forested peat plateaus and palsas only 0–20 mg CH_4 per m^2 per day are released (Bubier et al., 1995). These data suggest that major thaw of the areas of frozen peat would cause large increases in emissions of greenhouse gases from the Mackenzie valley peatlands. It should be noted that peat also has the ability to retard ground thaw and this may result in a lag in the response of peatlands to climate warming (Halsey et al., 1995).

CONCLUSIONS

Both the extent (area and volume) and the thermal state of peatlands in the Mackenzie valley are important for assessing potential ground behaviour in the case of climatic warming or other surface disturbances. Peatlands in areas of discontinuous permafrost are particularly significant as these areas are potentially most vulnerable to the effects of global warming (*see* Wright et al., 2000). Much is already known about the areal extent of bog and fen, but more field investigations are needed to accurately characterize the peat depths and ground temperatures associated with the different classes and forms of peatlands in boreal and subarctic regions.

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Postglacial vegetation and climate

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Abstract: The past climate and vegetation of the Mackenzie valley has been reconstructed for the past 13 000 radiocarbon years using fossil-pollen records and studies of tree rings. The earliest vegetation cover in much of the valley was sparse and dominated by small shrubs and herbs. It is possible that aspen and/or poplar trees were also present. By about 10 000 BP birch tundra was dominant throughout the valley. Between 10 000 and 9000 forests of white and black spruce were established as far north as the Tuktoyaktuk Peninsula. The forest retreated slightly in the Mackenzie Delta region and in the mountains after 5000 years ago. Tree-ring records show that the during portions of the nineteenth century, summer temperatures in the valley were perhaps 1 to 2°C colder than they are today. In the twentieth century there has been some increased establishment of spruce at higher elevations in the mountains.

Résumé : On a reconstitué le climat et la végétation de la vallée du Mackenzie au cours des 13 000 dernières années (âges au radiocarbone) en utilisant les renseignements fournis par le pollen fossile et les cernes de croissance des arbres. La première végétation à pousser dans la grande partie de la vallée était éparse et surtout composée de petits arbustes et de plantes herbacées. Des trembles ou des peupliers y poussaient peut-être également. Aux environs de 10 000 ans, la végétation dominante de la vallée était celle d'une toundra à bouleau. Entre 10 000 et 9 000 ans, des forêts à épinette blanche et à épinette noire se sont implantées aussi loin vers le nord que la péninsule de Tuktoyaktuk. Après 5 000 ans, les forêts ont reculé légèrement jusque dans le delta du Mackenzie et dans les montagnes. Les cernes de croissance des arbres indiquent que durant certaines périodes du 19e siècle, les températures estivales étaient peut-être plus froides de 1 à 2 °C que celles d'aujourd'hui. Au 20e siècle, on a observé une certaine augmentation des populations d'épinettes à plus haute altitude dans les montagnes.

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CLIMATE AND VEGETATION

The vegetation and climate of the Mackenzie valley have changed in the past and will change in the future. For example, approximately 18 000 BP (all data given here when discussing pollen records are radiocarbon years before 1950. Such ages do not correspond directly to calendar years) the entire Mackenzie valley was covered by glacial ice. Some of the ice came from valley glaciers and ice fields in the Mackenzie Mountains and some was part of the great Laurentide Ice Sheet that was centred to the east (*see* Dyke and Prest, 1987; Duk-Rodkin and Hughes, 1991; Duk-Rodkin and Lemmen, 2000). The glacial climate was extremely cold and vegetation was absent from the valley. By 13 000 years ago, the ice had begun to melt and the landscape of the Mackenzie valley became available for colonization by plants. A warming in climate had caused the melting of the ice. This warming and other climatic variations have caused changes in the vegetation of the valley over the past 13 000 years. For instance, 6000 years ago, summers were warmer and forests grew on the Tuktoyaktuk Peninsula, where only tundra exists today. In the near future, warming of the climate due to the 'greenhouse effect' may cause significant changes in the environment and vegetation of the Mackenzie valley. By studying how plants colonized the Mackenzie Basin, and how past changes in climate have affected vegetation, we can gain important insights into how the present environments of the Mackenzie valley evolved and how future climate changes, both natural and human, may affect the valley.

The changes in vegetation and climate that have occurred in the Mackenzie valley over the postglacial period (past 10 000 years) have been reconstructed using two lines of evidence. The first line of evidence is the record that comes from fossil pollen grains produced by plants in the past and deposited by wind and water in lake sediments and peat. The second is the analysis of tree-rings from dead and living trees. Both of these approaches are outlined below and the insights they provide on the past vegetation and climate of the Mackenzie valley are presented.

FOSSIL-POLLEN ANALYSIS

Common boreal plants such as white and black spruce (*Picea glauca* and *P. mariana*), jack pine (*Pinus banksiana*), birch (*Betula*), alder (*Alnus*), willow (*Salix*), grass (Poaceae), and sedge (Cyperaceae) produce large quantities of wind-borne pollen. The pollen grains are about the same size as silt (10–150 μm in diameter) and are easily blown away from the parent plant. Much of this pollen never fulfills its reproductive function and is deposited in lakes and bogs. Although the living interior of the grains decays rapidly, the exterior walls of the pollen grains are made up of a polymer called sporopollenin that is resistant to decay in settings where oxygen is limited, such as lake bottoms and damp, fast-growing peatlands. In these settings, pollen can be preserved for millions of years (Fig. 1).

The sediments on the bottom of a lake or in a peat bog collect over time, so that the oldest sediments are at the base and the youngest are at the surface. By studying the distribution of fossil pollen grains stratigraphically from the base of the sediments to the top, it is possible to reconstruct vegetation changes from the initial formation of the lake or bog until the present day. To undertake this research, cores are taken from the lake sediments and peat. In the laboratory, subsamples of a few millilitres are taken from the cores and chemical treatments are used to separate the pollen grains from the other organic and inorganic components of the sediments and peats. As many as 50 000 pollen grains can be recovered from 1 mL of sediment from small lakes in the Mackenzie valley.

The pollen grains are identified using a microscope at a magnification of about 400 x. Usually some 300 to 1000 grains are identified per sample. For some plants it is possible to identify the precise species that produced the pollen. For example, one can often tell apart the pollen of white spruce from black spruce. In the case of other plants, such as willows, one can only tell the genus. For some plants, such as grasses, it is generally only possible to identify the pollen as coming from the grass family (Poaceae). There are many plants, such as tamarack (*Larix laricina*), that do not produce large quantities of pollen. Other plants such as aspen and balsam

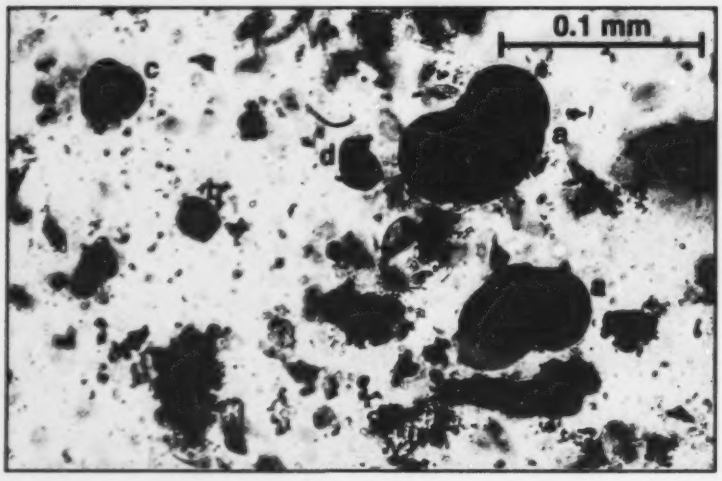

Figure 1.

A photomicrograph of fossil pollen grains from a lake near Yellowknife. The pollen includes: a, spruce, b, birch, c, heath, d, sedge. The pollen dates from about 5000 BP. Photograph by G.M. MacDonald. GSC 2000-027 (100 μm = 0.1 mm)

poplar (*Populus tremuloides* and *P. balsamifera*) produce pollen that is not well preserved in sediments. It is harder to reconstruct the history of these species on the basis of pollen. Despite this, the general character of the vegetation surrounding a lake or bog can usually be reconstructed from the pollen deposited in the sediments. Small lakes (1 ha or less) tend to receive pollen locally. Larger lakes receive a greater proportion of pollen from outside the immediate area and present a more regional picture of vegetation change.

Scientists studying fossil pollen pay careful attention to the modern relationship between vegetation and climate. If aspects of modern vegetation can be related to climate, such as the northern treeline, which appears to correspond with the northernmost locations where the mean July temperature is about 10°C, then past climate conditions can be inferred from the past distribution of vegetation. A past northward shift in treeline might be apparent in fossil-pollen records and interpreted as a period when July temperatures were warmer than at present. As many lakes in the Mackenzie valley formed at the end of the last ice age, it is possible to reconstruct the entire postglacial history of vegetation and climate in the valley.

Ages are assigned to the stratigraphic records of vegetation change by radiocarbon dating organic material in the sediments. Radioactive carbon (^{14}C) is formed naturally in the stratosphere and makes up a very small proportion of the total carbon found in the atmosphere and on Earth. Plants incorporate ^{14}C during photosynthesis and it is then passed up the food chain through animals that feed on plants. All living organisms contain a small amount of radioactive carbon. When a plant or animal dies, the ^{14}C undergoes radioactive decay and becomes nitrogen (^{14}N). Half of the radioactive carbon decays to nitrogen every 5730 years. By measuring the amount of radioactive carbon left in the remains of a plant or animal, scientists can determine how long ago the organism died. By convention, such ages are reported as radiocarbon years Before Present (BP) with 'Present' taken to be AD 1950. Radiocarbon years do not exactly equal calendar years. For the last few thousand years the difference between radiocarbon and calendar years is small. However, this difference increases back in time and a radiocarbon date of 10 000 BP represents a calendar date closer to 12 000 years ago.

A latitudinal transect of fossil-pollen records from lakes and bogs is available from Mackenzie valley (Fig. 2). This transect can be used to reconstruct vegetation change since the end of the last glacial period. In the extreme north, a fossil-pollen record from Sleet Lake in the vicinity of Tuktoyaktuk presents a record of 13 000 years of vegetation change (Spear, 1983, 1993; Fig. 2a). At 13 000 BP the region was newly free of glacial ice and supported tundra vegetation dominated by shrub birch. This is evidenced in the pollen record by the large amounts of birch pollen deposited at that time and very low amounts of tree pollen. Between 10 000 and 9000 BP there was an increase in spruce pollen. The abundance of spruce pollen remained high until about 7000 BP and then declined until 4000 BP. These changes in spruce-pollen abundance suggest that between 9000 and 7000 BP the region near Tuktoyaktuk was likely forested (Spear, 1983, 1993; Ritchie, 1984, 1987). This forest thinned and retreated between 7000 and 4000 BP. The present mean July temperature (an indication of how warm the summer is) on the Tuktoyaktuk Peninsula is approximately 7–9°C. The presence of spruce forest in this region between 9000 and 7000 BP suggests that the mean summer temperatures at that time may have been approximately 1–3°C warmer than those of today. In addition, there is evidence that the treeline was slightly higher in elevation and high-elevation forests were denser in the Mackenzie Mountains during this warm period (MacDonald, 1983; Szeicz et al., 1995) Needles of spruce have been found in peat deposits that lie above the modern treeline near the headwaters of the Natla River in the Mackenzie–Selwyn mountains (Fig. 2). The modern tundra that makes up the vegetation of the Tuktoyaktuk Peninsula today was established after 5000–4000 BP (Fig. 2a). In the pollen record this tundra is evidenced by large amounts of birch (shrub birch), alder, heath, grass, and sedge pollen (Ritchie, 1984, 1987; Anderson et al., 1992). It is known that the orbital geometry of the Earth was different in the past, and summer insolation in the high northern latitudes was about 4% higher than today between 11 000 and 6000 BP. This increased insolation would have caused summers to be warmer and could explain this northward shift of treeline in the Mackenzie valley and the upward shift of the treeline in the mountains (Ritchie, 1987; Spear, 1993; Szeicz et al., 1995).

Fossil pollen from Lac Mélèze and Lac Demain located further south in the valley near Norman Wells (MacDonald, 1987; Fig. 2b) and Fort Simpson (MacDonald, 1987; Fig. 2c) present a different pattern of vegetation history. The southern and central valley was deglaciated later than the Tuktoyaktuk region and the pollen records do not begin until 11 000 to 10 000 BP. Both of these sites were first occupied by a herb- and grass-dominated vegetation. This is recorded in the pollen records by large percentages of sage (*Artemisia*) and grass pollen. There is some pollen from aspen or balsam poplar in the sediments of this age, so it is possible that these trees were present. Aspen/poplar pollen is also present in the sediments deposited in Sleet Lake at this time (MacDonald, 1987; Spear, 1993; Fig. 2a). Perhaps the climate of the central and southern Mackenzie valley was dry rather than extremely cold when this initial vegetation was established. The herb, grass, and poplar vegetation was rapidly replaced by shrub birch. The pollen records show that the birch-dominated vegetation was in turn rapidly replaced by spruce forest between 10 000 and 9000 BP. Dense spruce forest has persisted at these sites from about that time until the present (Ritchie, 1984, 1987; Ritchie and MacDonald, 1986; MacDonald, 1987; MacDonald and McLeod, 1996).

At about 7000–6000 BP there was an increase in alder pollen in all sites along the Mackenzie valley (Fig. 2a–2c). Alder is typically associated with damp acidic soils and is often found growing around lake margins and stream channels. The near-synchronous increase in alder throughout the valley suggests an increase in moisture at 7000–6000 BP. This may have been due to decreasing summer warmth, increasing precipitation, or some combination of these factors (MacDonald, 1987).

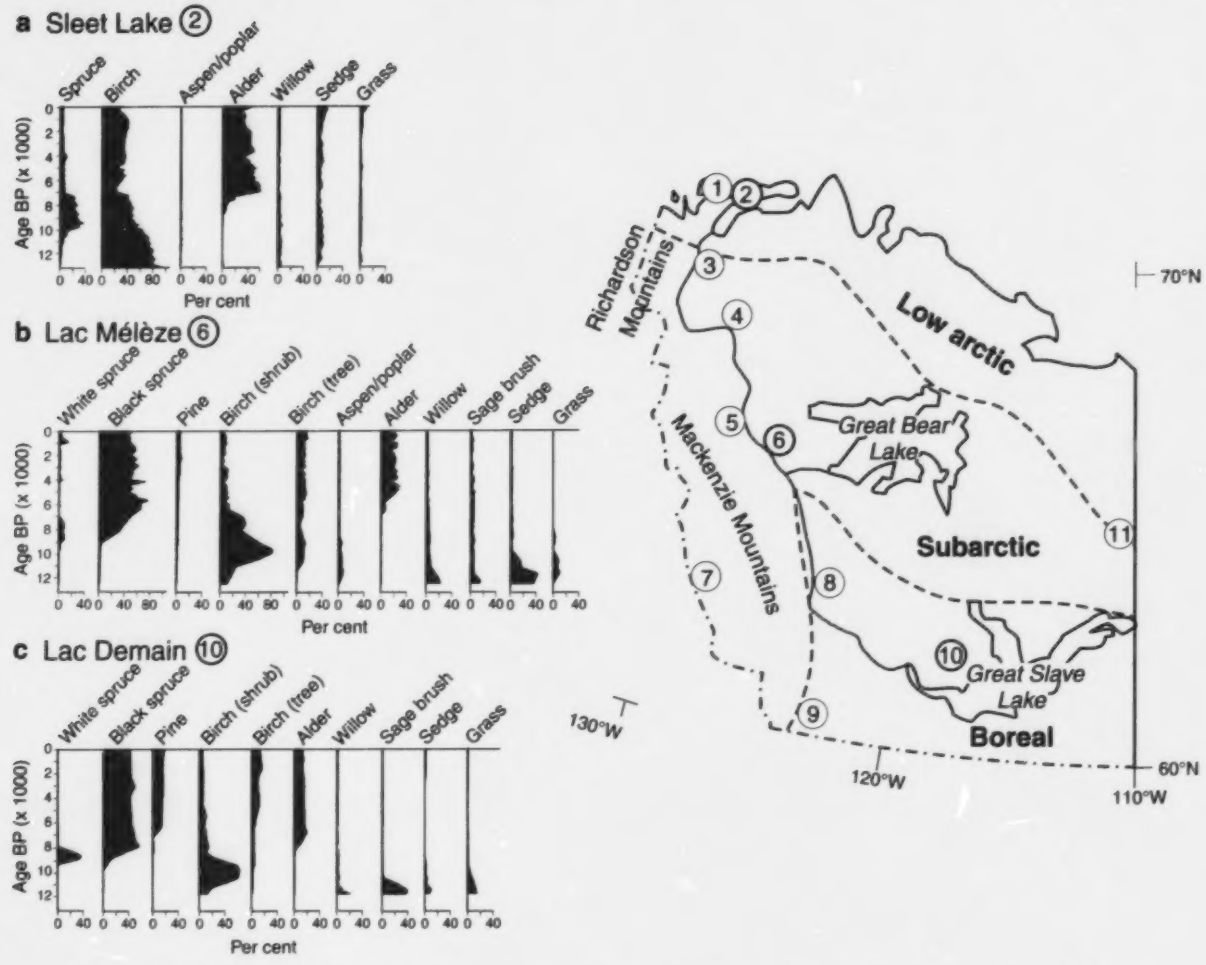

Figure 2. *The location of fossil pollen sites along the Mackenzie valley and fossil pollen diagrams from* ***a****) Sleet Lake (2),* ***b****) Lac Mélèze (6), and* ***c****) Lac Demain (10).*

Peatlands are an important part of the landscape of the Mackenzie Valley today and have been in the past (Zoltai and Tarnocai, 1975). The spores of *Sphagnum* peat moss are deposited and preserved in lakes and bogs along with pollen. The amount of these spores in sediment deposits increased to modern abundances between about 7000 and 4000 BP in the Mackenzie valley (MacDonald, 1987; MacDonald and McLeod, 1996; Fig. 3). This suggests that the extensive peatlands that typify the region today were established at that time. Peatlands are generally associated with cool and moist climates. A decrease in summer warmth and increasing moisture after 6000 BP may have aided the development of extensive peatlands in the Mackenzie valley (MacDonald, 1987).

The last important change was the arrival of jack pine in the southern and central Mackenzie valley at about 4500 BP. Both spruce and jack pine spread to the Mackenzie valley from the south (Ritchie and MacDonald, 1986; MacDonald, 1987; MacDonald and MacLeod, 1996). Spruce spread rapidly up the valley and was present from the Alberta–Northwest Territories border to Tuktoyaktuk by 9000 BP. The rapid spread of spruce may have been due to the northward transport of seed by northwesterly blowing winds coming from the remaining Laurentide Ice Sheet (Ritchie and MacDonald, 1986). In addition, seeds, and even trees and soil dislodged by floods may have been transported northward by the Mackenzie River. The reasons for the slower spread of pine compared to spruce are unclear. It may have been that the glacial refuge of jack pine was located further to the southeast than that of spruce. In addition, jack pine cannot tolerate shade and may have had a hard time establishing once spruce forests were dominant in the region. Finally, a changing climate after 6000 BP, perhaps to moister conditions, may have helped the spread of jack pine.

Taken together, the fossil-pollen evidence from sites along the Mackenzie valley suggest that at 13 000 BP the northern end of the valley supported a shrub-birch-dominated vegetation (Fig. 4a). By 10 000 BP, the southern and central portions of the valley were ice-free and supported a herb-,

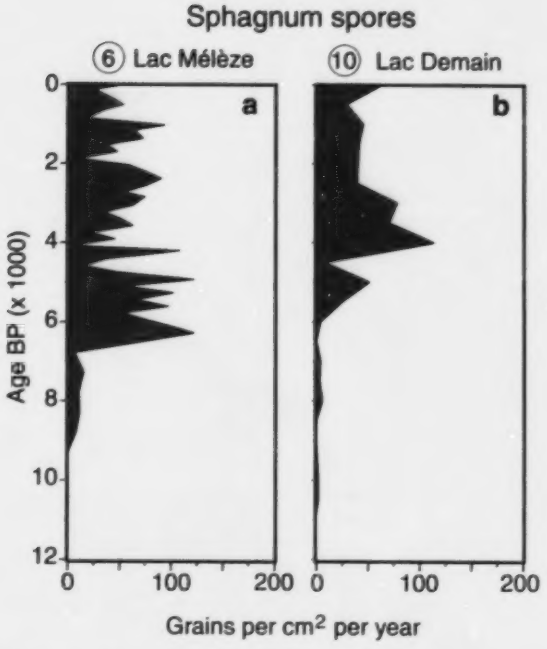

Figure 3. *The accumulation rates of peat-moss spores (Sphagnum) in lake sediments at* ***a****) Lac Mélèze (6), and* ***b****) Lac Demain (10). The large increase at around 6000 BP likely represents the establishment of widespread peatlands in the Mackenzie valley.*

shrub-, and poplar-dominated vegetation (Fig. 4b). Warm, and relatively dry summers occurred between 10 000 and about 6000 BP. The initial herb, birch, and poplar vegetation was rapidly replaced by spruce forest between 10 000 and 9000 BP. Spruce trees grew well north of their modern range between 10 000 and 5000 BP (Fig. 4c). Between 6000 and 4000 BP, the spruce treeline retreated southward, alder and peatlands became important components of the vegetation, and jack pine appeared in the valley. By about 4000 BP, the modern vegetation of the region was established. Since that time there has been little in the way of major vegetation or climatic changes. However, it is likely that smaller scale changes in vegetation have occurred and some of these will be examined below.

ANCIENT WOOD AND TREE-RING ANALYSIS

The study of ancient wood and tree rings provides another line of evidence for reconstructing past vegetation and climate in the Mackenzie valley. In the cold and relatively dry conditions of the Arctic, wood may be preserved for thousands of years following the death of a tree. Several stumps of white spruce and tamarack have been found on the tundra of the Tuktoyaktuk Peninsula (Spear, 1983; Ritchie, 1984). These stumps date to a time when forests grew on the peninsula. Radiocarbon dates from the stumps range from 7500 to 4900 BP. The presence of these stumps, and the radiocarbon ages obtained for them, confirm the evidence from the fossil-pollen records of forest development on the Tuktoyaktuk Peninsula in the past.

Trees such as spruce produce distinctive concentric growth rings during each summer. The rings are made up of couplets of light 'early wood' that forms at the start of the

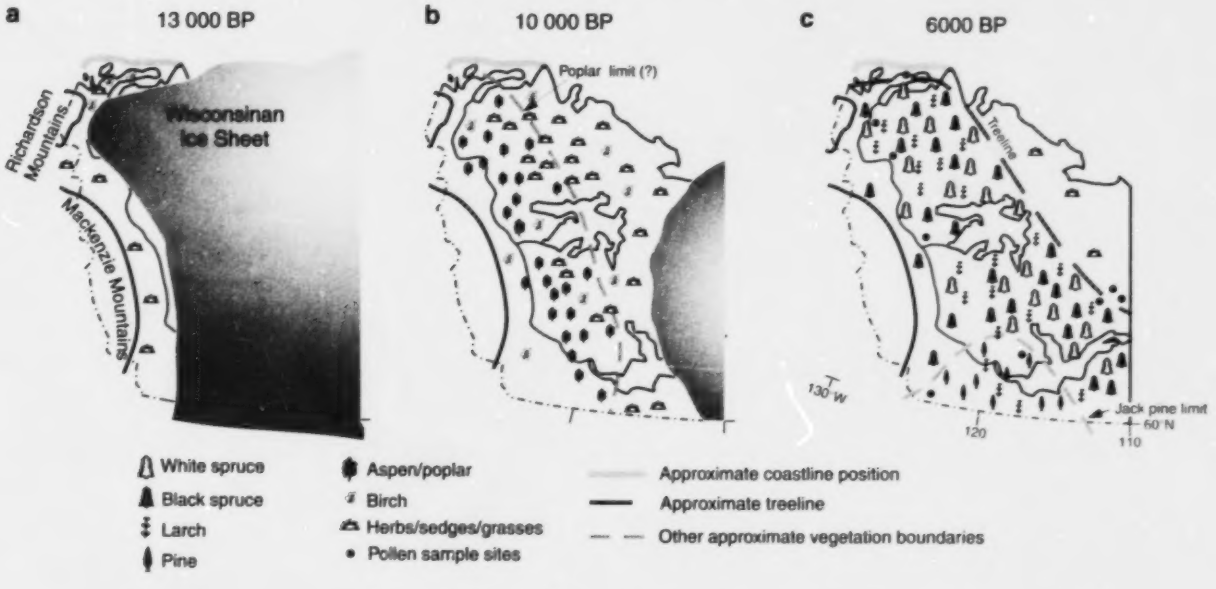

Figure 4. *Summary maps of vegetation in the Mackenzie valley at* ***a****) 13 000 BP,* ***b****) 10 000 BP,* ***c****) 6000 BP.*

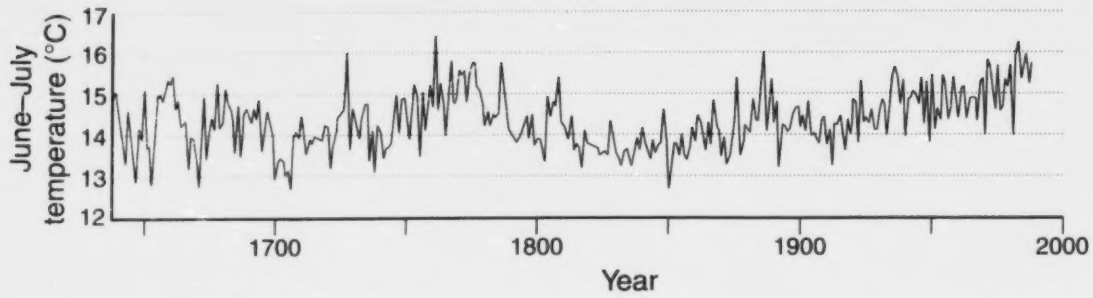

Figure 5. *Reconstruction of June–July temperatures in the Mackenzie valley region for the period AD 1638–1988.*

summer and darker 'late wood' that forms as growth slows down towards the middle of summer. The oldest rings are found at the centre (or pith) of the tree and the most recent near the bark. By counting these rings it is possible to determine the age of the tree. In general, trees growing under good conditions produce wide rings. For spruce trees near treeline in the northern Mackenzie, or in the mountains, the rings produced during years with warm summers are generally wide, while those produced during cool years are narrow (D'Arrigo and Jacoby, 1993; Szeicz and MacDonald, 1995a). Thus, the width of the rings can be used to gauge the general conditions under which the tree grew, and also the specific changes in weather from year to year.

The ring widths and growth rates of the stumps from the Tuktoyaktuk Peninsula have been measured and compared with the growth rates of living trees. The rate of growth of these ancient trees was found to be similar to that of the modern spruce trees growing near Inuvik. This suggests that summer climate near Tuktoyaktuk 6000 years ago was similar to the present climate at Inuvik, which is further south (Ritchie, 1984). This would represent a warming of July temperatures at Tuktoyaktuk of approximately 3°C. This estimate of summer warming corresponds well with estimates based on the fossil-pollen record as outlined above.

Tree-ring data collected from living white spruce trees, and spruce trees that have died recently, have been used to produce detailed records of summer temperature variations over the past 300 to 500 years in northwestern Canada and Alaska (D'Arrigo and Jacoby, 1993). In these studies, tree-ring records from a number of trees at several sites are mathematically combined to produce estimates of summer and annual temperatures. These reconstructions indicate that temperatures were low during the last century and have been generally increasing since that time. Detailed analysis of trees growing in the Mackenzie and Franklin mountains have recently been used to reconstruct July mean temperatures (Szeicz and MacDonald, 1995a). This reconstruction provides similar evidence of cold conditions during the period AD 1800 to 1880 with a general warming since that time (Fig. 5). During the very coldest period, at around AD 1850, temperatures in the Mackenzie valley were likely 2–3°C cooler than at present.

The cold conditions of the last century have been recognized at many sites in northern North America and Europe. The time period from AD 1600 to 1880 was generally cool and is often referred to as the Little Ice Age. Some have attributed the warming trend following AD 1880 with the greenhouse effect, caused by increasing level of anthropogenic carbon dioxide in the atmosphere that started with the Industrial Revolution. The warming since 1880 may, however, be a natural variation in climate (*see* Dyke and Brooks, 2000).

The impact on vegetation of these cold temperatures of the last century and the more recent warming has been studied at the treeline in the Mackenzie and Franklin mountains (Szeicz and MacDonald, 1995b). It has been found that many trees at the treeline died during the middle of the last century. Many of the trees which grow at these high-elevation sites today did not begin to grow until the start of warming in the 1880s. Thus, the treeline vegetation that we see today is actually a product of changing climate over the past 200 years.

ACKNOWLEDGMENTS

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Tree-ring evidence of recent climate changes in the Mackenzie Basin, Northwest Territories

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Bégin, C., Michaud, Y., and Archambault, S., 2000: Tree-ring evidence of recent climate changes in the Mackenzie Basin, Northwest Territories; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 65–77.*

Abstract: Three long-term climate reconstructions based on tree-ring analysis for the boreal and forest tundra biozones in the Mackenzie valley are presented. The northernmost site (Eskimo Lakes site) provides the longest tree ring chronology, covering the period of 1172–1991, while the boreal sites (Mountain River) span the last five hundred years. Even though sampled sites correspond to different environments and are separated by hundreds of kilometres, they show similarities in long-term climatic trends. The period known as the "Medieval Warm Period" was follow by a long period of cold, dry conditions between 1330 and the early 1900s. This suppressed growth period, corresponding to the "Little Ice Age", was briefly interrupted by milder conditions during the first half of the sixteenth century. Finally, all tree-ring chronologies show a increasing growth at the beginning of the twentieth century.

Résumé : Trois reconstitutions climatiques à long terme fondées sur l'analyse des cernes de croissance des arbres dans les biozones de la forêt boréale et de la toundra dans la vallée du Mackenzie sont présentées. Le site le plus au nord (lacs Eskimo) livre la dendrochronologie la plus longue, qui s'étire sur la période de 1172 à 1991, tandis que les sites de la forêt boréale (rivière Mountain) renseignent sur les cinq derniers siècles. Même si les sites échantillonnés correspondent à des environnements différents et qu'ils sont séparés par des centaines de kilomètres, les tendances climatiques à long terme qu'ils indiquent affichent des ressemblances. La période chaude médiévale a été suivie par une longue période de conditions froides et sèches entre 1330 et le début du 20e siècle. Cette période de croissance ralentie, correspondant au «Petit âge glaciaire», a été interrompue brièvement par des intervalles à conditions plus douces durant la première moitié du 16e siècle. Enfin, toutes les données chronologiques basées sur les cernes de croissance des arbres révèlent une croissance accrue au début du 20e siècle.

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INTRODUCTION

The Mackenzie valley encompasses different vegetation zones, ranging from boreal forest to shrub tundra. During the Holocene, the distribution of vegetation in the Mackenzie valley has undergone important modifications in response to major climatic changes. Such modifications occurred because the species present and tree growth in high latitudes are strongly influenced by temperature regime. In treeline environments, moisture is normally adequate and tree growth is highly dependent on summer temperatures (growing season). Tree rings offer the greatest potential for reconstruction of climate because of the large geographical distribution of suitable sites, the annual resolution, and the relatively long time span of records. Moreover, several studies have demonstrated that boreal species in marginal environments, especially at the northern treeline, retain a strong temperature signal in their ring-width records (Jacoby and Cook, 1981; Jacoby and D'Arrigo, 1989; D'Arrigo and Jacoby, 1992).

This paper presents climatic interpretations from tree-ring chronologies obtained within the two most important vegetation zones in the Mackenzie valley: the boreal forest and the forest tundra. The first chronology is from Eskimo Lakes area near the Tuktoyaktuk Peninsula (Fig. 1a) and reveals climatic changes over the past 800 years near the treeline in the high subarctic area. Two other chronologies come from the Mountain River area (Fig. 1b) located in the upper section of the boreal forest. One of these is representative of sheltered conditions (forest chronology) while the other represents exposed conditions (krummholz chronology). In all cases, the presence of subfossil material (dead stems lying on the surface or still in living position) allows chronologies to be extended substantially beyond those given by living trees.

DENDROCHRONOLOGY AND DENDROCLIMATIC TECHNIQUES

Climate interpretations from tree rings can be extended into the past, well beyond the age of living trees, by crossdating. Individual trees are sliced or bored, and the width of the rings measured, usually with the aid of a binocular microscope. A graph of ring width versus age in years gives a tree-ring series. Crossdating requires a search for distinctive patterns of wide or narrow rings and matching these, first between living and dead trees, then between progressively older dead trees. This procedure gives a ring-width chronology with an exact year of formation for each ring. It is beneficial to have

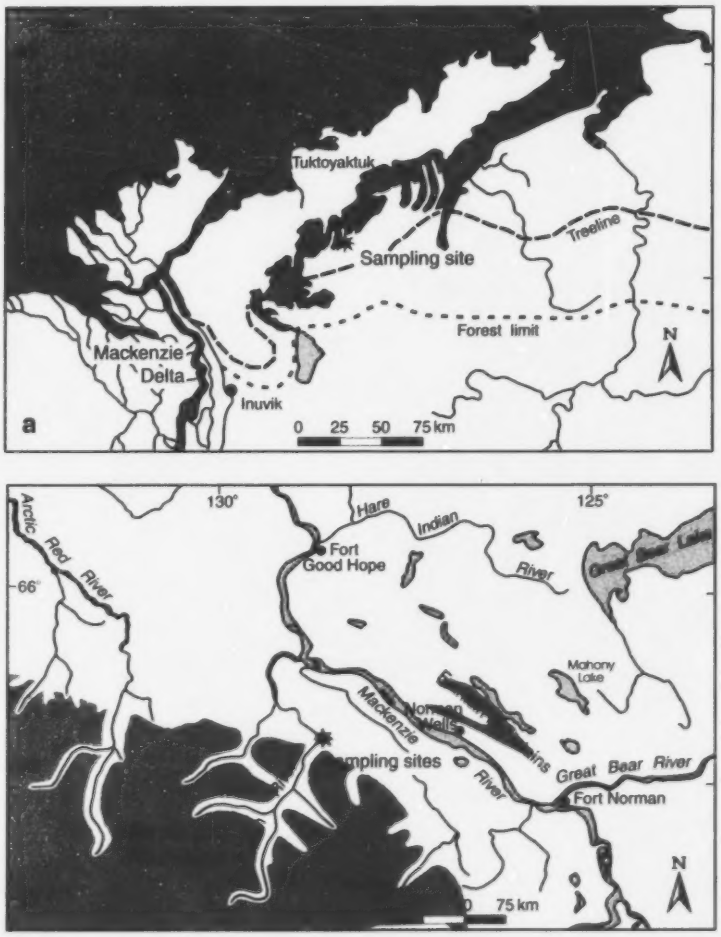


Figure 1.

Location of sampling sites: ***a)*** *Eskimo Lakes site and* ***b)*** *Mountain River sites.*

several trees of overlapping age to assist with the recognition of false rings, missing rings, or frost rings formed in response to extreme conditions.

The raw ring-width sequences developed in this study are composed of several series, each of which may show influences not directly linked to climate. These influences include a tree's biological growth trend, whereby ring widths generally tend to decrease with age, or the effects of natural disturbances such as landslides. At the Mountain River site, long-term growth of trees was influenced by slow eolian sedimentation. These trends are removed from each individual series by a standardization process.

The standardization process is done by fitting a smooth curve to each tree-ring series and dividing each ring width by the corresponding value given by the curve. The overlapping parts of these transformed series are averaged and used to produce a 'standard chronology'. The standard chronology emphasizes short-term variation over the life of individual trees used to construct the chronology, but eliminates information on longer term climatic variation. To obtain longer term trends present in tree-ring series, each ring's width was divided by the mean ring width for that series to produce a normalized series. These series show trends spanning the full average life of trees and are used to construct an 'absolute chronology' (Archambault and Bergeron, 1992).

Under some conditions, ring width characteristics of a given year are strongly influenced by the growth of previous years (biological persistence effect). Therefore, when an attempt is made to calibrate a tree-ring series with an instrumented climate record (e.g. measured temperatures at a meteorological station), the influence of climate factors may be obscured by the influence of biological factors. To determine if this is the case, each ring of a series can be compared with the previous year's ring. This is referred to as autocorrelation. If the autocorrelation is high, then biological effects are strong. Once recognized, they can be removed to produce a 'residual chronology'. This type of chronology is most valuable for determining the correlation of ring width with measured climatic variables.

The relationship between tree-ring chronologies and monthly climate variables was examined by correlating ring-width indices with monthly mean temperature and total precipitation for each month from April of the year before ring formation through August of the year of ring formation for the closest meteorological station (Inuvik station for Eskimo Lakes chronology and Norman Wells station for Mountain River chronologies). This range of climatic variables was chosen because it includes two complete growing seasons and allows the importance of factors acting during the year preceding ring formation to be assessed (i.e. bud formation, foliage protection, moisture storage). Reduced vegetative structure caused by severe conditions during the previous fall and winter (bud and needle desiccation and/or abrasion) will have a negative effect on photosynthetic activity and, consequently, on tree growth. Finally, because ring width rarely depends on a single climatic variable (Earle et al., 1994), various combinations of seasonal climatic variables were also investigated to better explain growth variation.

When the relationship between tree growth and monthly climate variables is sufficiently strong (r>0.5–0.6) to suggest a statistical link between a specific variable and ring width (Fritts, 1976; Garfinkel and Brubaker, 1980), this regression can be used as a basis for reconstructing climate variations before the period of climate records. This is true for the Eskimo Lakes chronology where tree growth was strongly correlated with mean July temperature. Since the data set in this case consists of a single chronology derived from one measured variable (total ring width) the model consisted of a simple linear regression. Even when there is no clear statistical connection between a specific climate variable and ring width, past environmental conditions, including climate factors, can be estimated directly from longer term tree-growth patterns.

Trees at the Eskimo Lake and Mountain River sites were sampled between 1990 and 1992. Because most of the sampled trees come from exposed sites where radial growth series may contain several growth anomalies (missing rings, frost rings, etc.) reflecting extreme growing conditions, complete cross-sections were taken as close to ground level as possible.

TREELINE SITE — ESKIMO LAKES CHRONOLOGY

Study site

This site is located on the south shore of Eskimo Lakes (69°08′N, 132°13′W) about 105 km northeast of Inuvik and 48 km southeast of Tuktoyaktuk (Fig. 1a). It is representative of the arctic treeline in the lower Mackenzie valley and lies within a continental subarctic climate. Meteorological data recorded since 1957 at the Inuvik weather station show a mean annual air temperature of -96°C (Fig. 2), and a mean monthly air temperature for July of 14°C. On average, the degree-days of growth (>5.5°C) equal 654. Total annual precipitation is 256 mm, while total annual snowfall is 175 cm (55% of total precipitation). Summers are characterized by frequent incursions of Pacific air masses (30–40% of days) in the form of rain-producing cyclones (Ritchie, 1984). The site is underlain by continuous permafrost.

Tree-ring samples were collected from white spruce trees (*Picea glauca* (*Moench*) *Voss*) that lie in the northernmost part of the forest tundra, between the forest limit and the tree-growth limit. Small forest stands (up to 7 m tall) surrounded by low shrubs characterize the vegetal landscape (Fig. 3). No evidence of fire (charcoal, snags, etc.) was found on the site, suggesting that these small stands evolved naturally without fire disturbance for several centuries or even millennia. This old-growth treeline vegetation is composed of living and dead white spruce stems of all ages and of different shrubby or normal growth forms. Tree-ring dating of the dead population, still in growth position or lying on the surface, indicates that subfossil stems can remain undecomposed for up to 500 years depending on site conditions. For these reasons, this site offers a high potential for a reconstruction of climatic history at the treeline.

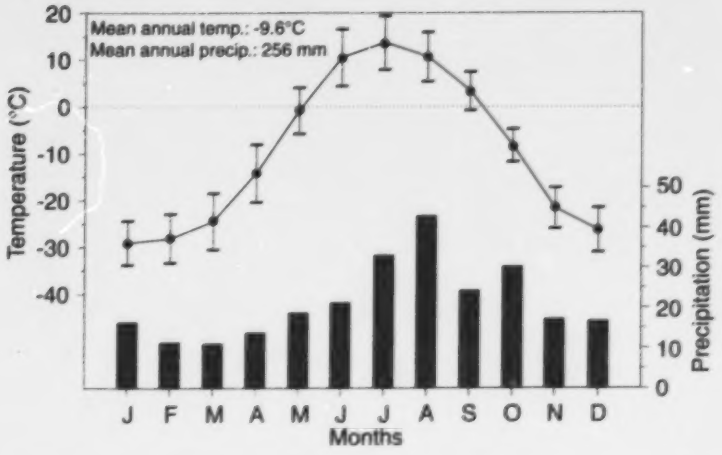


Figure 2.

Mean monthly temperature and total precipitation, Inuvik weather station (1957–1989). Error bars indicate standard deviation for temperature (from Environment Canada, 1993).

Figure 3. *Old-growth white spruce stand at the treeline on the south shore of Eskimo Lakes (Eskimo Lakes sampling site) which is characterized by the presence of living and subfossil individuals. Photograph by C. Bégin. GSC 2000-028A*

Dendrochronology

Fifty-four radii from 28 trees (26 radii from living trees and 28 radii from dead trees) were crossdated and averaged to produce the Eskimo Lakes chronology, covering the 820 years from AD 1172 to 1991 (Fig. 4). The main difficulty encountered was the extreme narrowness of the annual rings, the mean ring width being only 0.18 mm. Nevertheless, of the 12 909 annual rings crossdated, only seven locally absent rings were identified (0.05%). The mean correlation between the 54 series and the master chronology was 0.524, a figure indicating strong common variance and excellent crossdating between all samples.

Variations in absolute ring widths, especially in the smoothed form (Fig. 4b), show two general periods of increasing growth, separated by a period of slow growth. The first accelerated period of growth occurred between AD 1172 and 1335, when growth remained far above the mean for each year except for 1248 to 1275, when tree growth experienced a severe depression; however, the portion of the chronology between 1172 and 1265 must be interpreted cautiously since it is represented by only two series. This period of strong tree growth was then followed by a lengthy period (1336–1903) when growth was mostly below the mean. During these 567 years of slow growth, the fifteenth and sixteenth centuries showed slight improvement in growth within the interval 1507 to 1512. The following three centuries (seventeenth to nineteenth) represent the most difficult period for tree growth; the lowest ring-width values were recorded in 1652, 1685, 1707, 1721, and 1827. Since 1904, tree growth has begun to increase rapidly, reaching a maximum in 1948. Since then, tree growth has decreased slightly, yet remained above the mean value.

Climate reconstruction

Of the 23 variables analyzed in the correlation function procedure, only July mean temperature ($r = 0.559$) and previous November precipitation ($r = 0.417$) were significantly correlated with annual growth expressed as the residual chronology (Fig. 5). Combinations of monthly temperatures gave lower correlation values than for July mean temperature alone. The standardized chronology was also subjected to the correlation analysis to check if it was influenced by the same variables. The response was quite similar for both chronologies, although correlations with the standardized chronology were not as strong as with the residual chronology. The positive correlation between growth and July temperatures has been observed in different subarctic areas (e.g. Garfinkel and Brubaker, 1980; Jacoby and Cook, 1981). It suggests that high temperatures during the growing season enhance tree growth by increasing soil temperature, irradiance, and net photosynthesis. The positive response of growth to previous November precipitation probably reflects the protective role of snow cover against winter desiccation and foliage abrasion by wind.

The mean July temperature reconstruction is simply a rescaling of the original residual chronology (Fig. 6a). A smoothed version of the reconstruction expressed as deviation from the 1957–1989 Inuvik mean (Fig. 6b), shows the twentieth-century average July temperature to be 0.05°C warmer than present temperature and at least 0.6°C warmer than previous centuries. This is probably an underestimation of the actual temperature increase since it is obtained from the residual chronology. The standard chronology is also

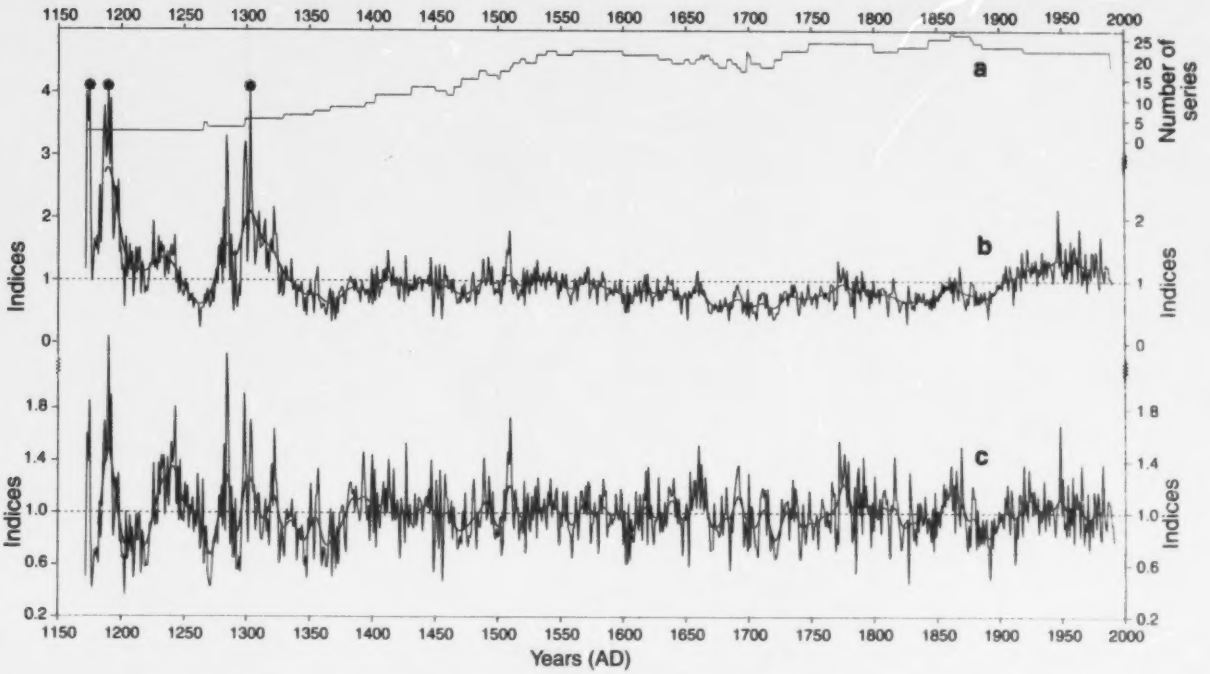


Figure 4. *Tree-ring chronologies;* ***a)*** *the number of series included in each portion of the chronology;* ***b)*** *absolute ring width; and* ***c)*** *standardized chronologies from Eskimo Lakes (treeline) site. Smoothed lines are 15 year running means. Indices above 4.0 (*) are truncated.*

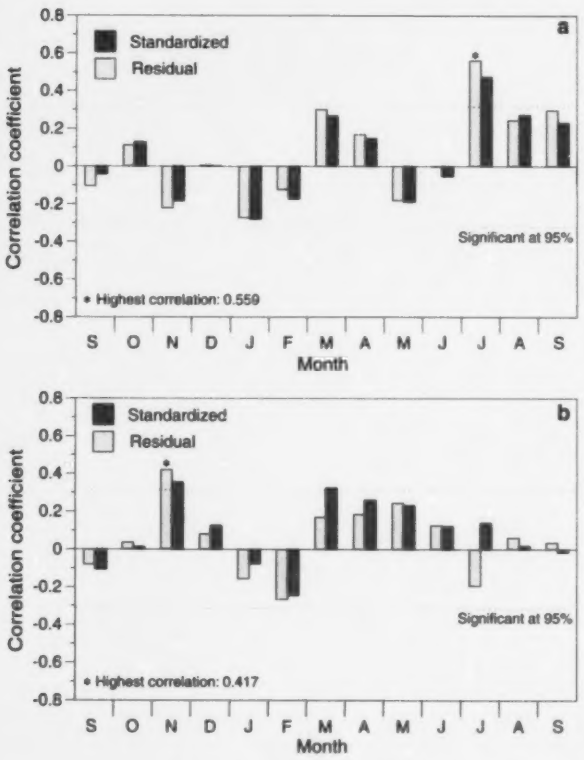


Figure 5.

Correlation of tree growth with ***a)*** *monthly mean temperature and* ***b)*** *total monthly precipitation at Inuvik (1957–1989) for the standardized and residual Eskimo Lakes treeline chronology.*

positively related to July mean temperature, although at a lower level of confidence. A regression with the standard chronology suggests a 1.5°C increase in mean July temperature relative to the seventeenth to nineteenth centuries. This correlates very well ($r = 0.519$, $p < 0.0001$) with a 374-year reconstruction of northern hemisphere annual temperatures (D'Arrigo and Jacoby, 1992) developed using boreal treeline sites. The colder July temperatures prior to the twentieth century correspond to the period known as the Little Ice Age (Grove, 1988). Our tree-ring data indicate that this period would have extended from 1336 to 1903 at the treeline in the north Mackenzie valley. Except for some short intervals when July temperatures have been above the present value (particularly 1507 to 1512), the Little Ice Age July temperature was about 0.3°C colder than the present normal.

BOREAL SITES (FOREST AND KRUMMHOLZ) — MOUNTAIN RIVER CHRONOLOGIES

Study sites

A forest and a krummholz (stunted, gnarled growth often found at the treeline) site in the vicinity of Mountain River (65°14′N, 128°33′W) about 80 km west of Norman Wells (Fig. 1b) have also been studied. Climate characteristics have been provided by the Norman Wells weather station where complete records are available since 1943. Mean annual air temperature is -6.1°C (Fig. 7) with a large range in extreme daily temperatures varying from -43°C in January to 24°C in July. The average annual precipitation totals 321 mm. The study area is located in the upper Mackenzie section of the boreal forest where conifers locally occupy well drained terrains. Terraces and valleys have been colonized by forest in which white spruce (*Picea glauca* (*Moench*) *Voss*) is the dominant tree species, often found in association with gray alder

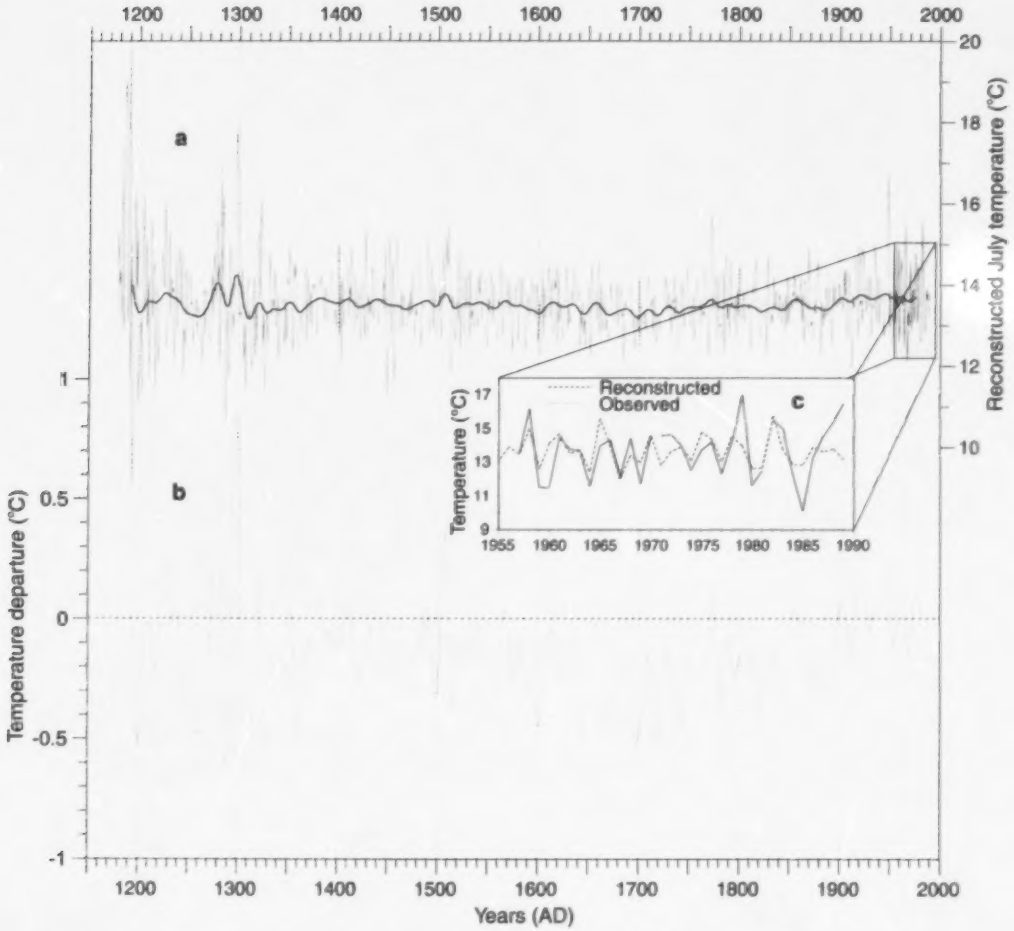


Figure 6. *Reconstructed July temperature at Inuvik for* ***a)*** *the period 1172–1991 and* ***b)*** *temperature departure from the 1957–1989 normal period. Comparison between the reconstructed (dashed) and actual (solid line) July temperature for the period 1955–1989 is shown in* ***c)****.*

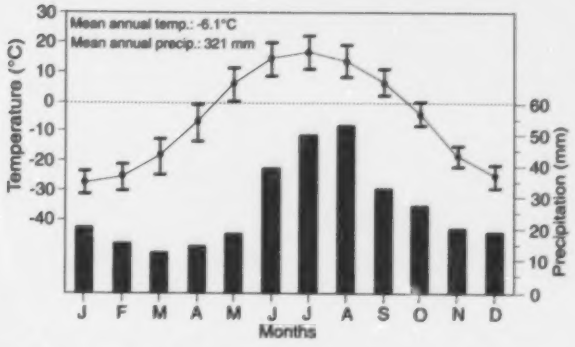


Figure 7. Mean monthly temperature and total precipitation, Norman Wells weather station (1944–1989). Error bars show standard deviation of temperature (from Environment Canada, 1993).

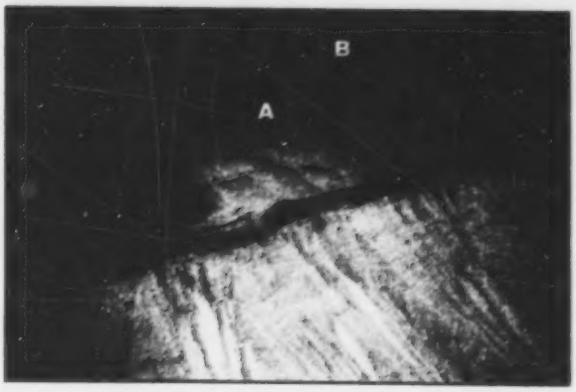


Figure 8. Oblique aerial photograph showing the white spruce forest 'A' and krummholz 'B' sampled stands at the edge of a glaciofluvial terrace along Mountain River. Photograph by C. Bégin. GSC 2000-028B

(*Alnus incana* (*L.*) *Moench*). In closed forest, various species of feather mosses cover the ground surface. Clonal populations of balsam poplar (*Populus balsamifera L.*) and various shrub species (e.g. *Eleagnus commutata Bernh.*) have colonized zones experiencing active alluvial or eolian deposition.

Both sites are located at the top of an 85 m high glaciofluvial terrace overlooking Mountain River (Fig. 8). The terrace surface is mostly occupied by peatland where stunted white spruce and larch (*Larix laricina* (*Du Roi*) *Koch.*) are growing. The terrace edge is affected by eolian deposition that generates, in some places, small cliff-top dunes (Bégin et al., 1995). Forest tree-ring samples were collected from this forest margin, while the krummholz tree-ring samples were collected from a stunted tree stand growing on a rocky ridge located behind the forest margin on a site experiencing a harsher growing environment.

Dendrochronology

Forest site

Several generations of trees that evolved in the cliff-top eolian environment (Fig. 8) were used to produce the Mountain River forest chronology. A total of 161 samples were taken from 1) fossil trees exposed in the deflation zone (n = 66), 2) both dead and living trees found on the dune slipface (n = 38), and 3) living trees in the periphery of the dune area (n = 57). Of all the samples collected, only 42 radii from 27 trees were crossdated and averaged to produce the forest chronology. Most samples were rejected to avoid the influence of sand deposition on growth patterns. Nevertheless, the resulting chronology covers a period of 471 years, from AD 1521 to 1991. Within the 11 130 annual rings crossdated, seven locally absent rings were identified (0.06%). The crossdating procedure was facilitated by the presence of reliable light rings, especially those from 1871, 1884, and 1959. The mean correlation between the 42 individual series and the master was 0.409.

Long-term environmental changes are clearly shown in the absolute ring-width chronology (Fig. 9b). Excluding the portion of the chronology prior to 1595, which is represented by less than three series, three distinct periods appear in the remaining 400 years. The interval between 1595 and 1777 is characterized by a slow growth where values oscillate around or slightly below the mean. After 1777, tree growth decreased substantially to reach its minimum value in 1809. It remained systematically below the mean until 1852. Within this period of suppressed growth, the interval from 1809 to 1813 represents the most difficult growing conditions. Since 1852, radial growth increased gradually until the end of the nineteenth century and reached maximum growth between 1899 and 1901. Radial growth decreased slightly between 1908 and 1948, but then decreased dramatically to the mean value of the present day.

Krummholz site

The sampled krummholz stand has similarities to the Eskimo Lakes site. It is an old-growth white spruce stand that has escaped natural fire for a long time and it is composed of living and dead trees rarely exceeding 3 m in height; however, this exposed vegetation is primarily characterized by the complex growth forms of trees. Multistem structures and localized foliage give the trees a bonsai-like appearance (Fig. 10).

A total of 65 stems (27 living and 38 subfossils) were sampled, but most of them were rejected after analysis because of the narrowness of their annual rings. In some cases, decades of rings were totally missing in the series. Twenty-seven radii from 17 trees were selected for the construction of the Mountain River krummholz chronology that covers 507 years, from AD 1485 to 1991. Out of the 8038 annual rings crossdated, only three locally absent rings were identified (0.036%). The mean ring width was 0.16 mm and the mean correlation between the 27 individual series and the master was relatively low, with r = 0.441.

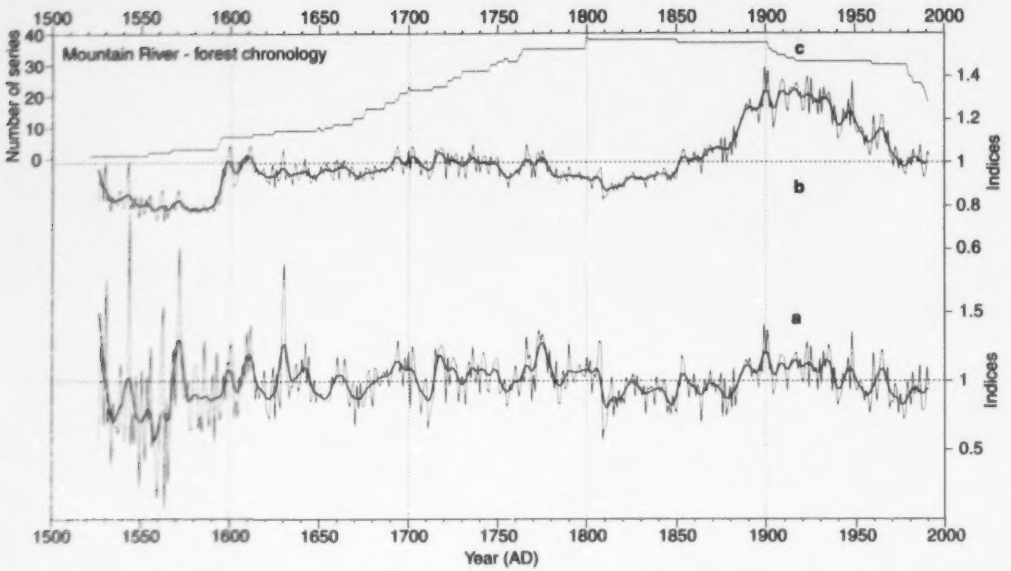


***Figure 9. a)** Standardized and **b)** absolute ring width chronologies from Mountain River (forest) site. Smoothed lines are 11 year running means. The number of series included in each portion of the chronology is shown in **c**).*

Environmental changes within the krummholz site are expressed in the absolute ring-width chronology (growth indices). Two periods of increased growth, separated by a long period of slow growth, appear in the long-term tree-ring pattern (Fig. 11b). The three periods are fortunately represented by enough series to be interpreted adequately (four tree-ring series occur before 1503 and six before 1533). The period between AD 1485 and 1573 is characterized by accelerated tree growth which remained far above the mean except for the intervals of 1500–1504 and 1534–1538, when growth decreased sharply. After 1573, tree growth decreased abruptly and remained mostly below the mean for the next three centuries, between 1574 and 1918. Within this lengthy period of slow growth, the intervals of 1602–1607, 1670–1675, 1832–1835, 1878–1884, and 1895–1898 were particularly critical for tree survival and the strongest negative deviations occurred in 1606, 1834, and 1895; however, this period was also characterized by a few intervals when growth exceeded the mean in response to better climatic conditions. Such intervals occurred between 1628 and 1637, 1772 and 1783, and 1801 and 1808. Finally, a period of improved growth conditions occurred during the twentieth century, between 1919 and today. Tree growth has increased progressively and reached a maximum value in 1955. Since that year, the growth has decreased.

CLIMATE INTERPRETATIONS

The two tree-growth curves show how growth conditions have fluctuated during the last centuries in the Mountain River area (Fig. 9, 11). The climatic significance of these variations can be estimated by examining the potential relationships between climatic factors and tree-ring values. The

***Figure 10.** General view of the old-growth, white spruce, krummholz stand at the Mountain River site (Mountain River krummholz sampling site) characterized by stunted subfossil and living trees. The growth form of the approximately 400-year-old living tree in the centre shows the erosional snow/air interface (1.05 m). Person at the left side gives the scale. Photograph by C. Bégin. GSC 2000-028C*

correlation of monthly mean temperature and monthly precipitation with forest and krummholz chronologies is presented in Figures 12 and 13, respectively. Both chronologies show a different response to climate.

In general, tree growth at the forest site shows a consistent negative response to summer temperature and a positive (although not statistically significant) response to precipitation in several months. In more detail, the mean temperature of the preceding June appears to be the most important climatic variable in terms of tree growth (r = -0.44); however,

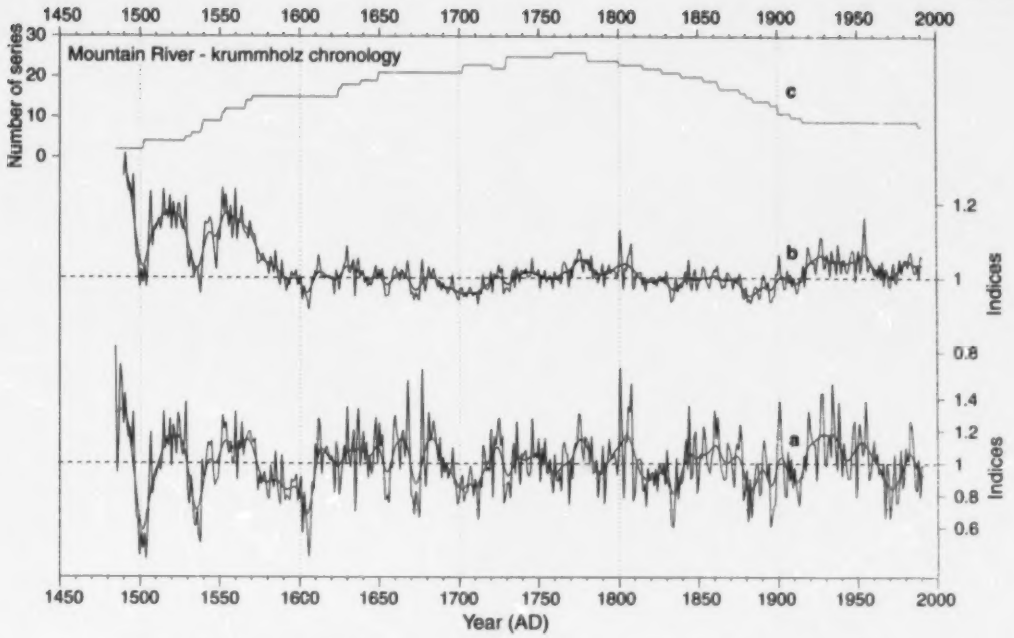
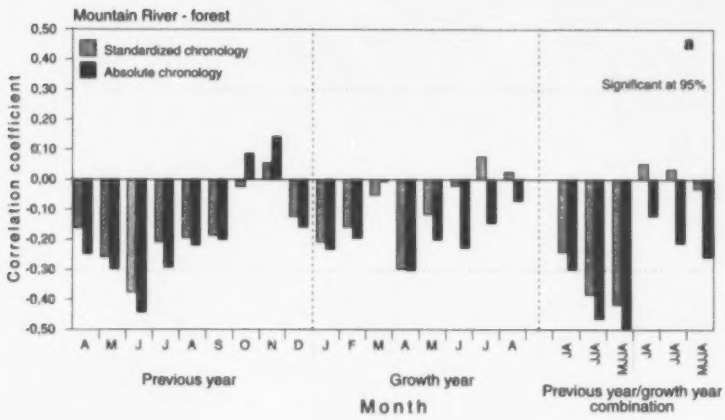


Figure 11. *a) Standardized and* **b)** *absolute ring width chronologies from Mountain River (krummholz) site. Smoothed curves are 15 year running means. The number of series included in each portion of the chronology is shown in* **c)**.



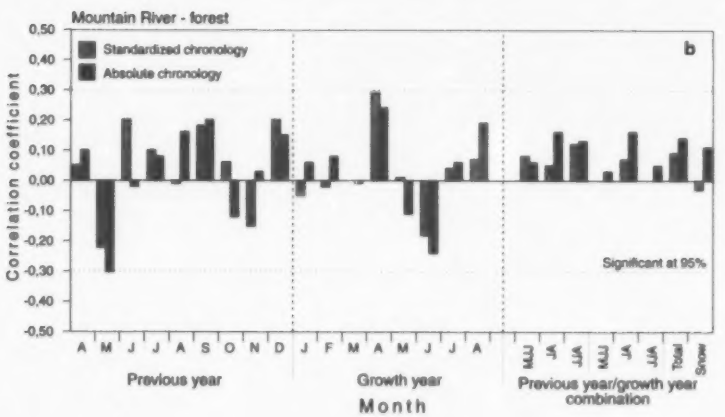


Figure 12.

Correlation of tree growth with **a)** *monthly mean temperature and* **b)** *total monthly precipitation at Norman Wells (1944–1989) for the standardized and absolute Mountain River forest chronology.*

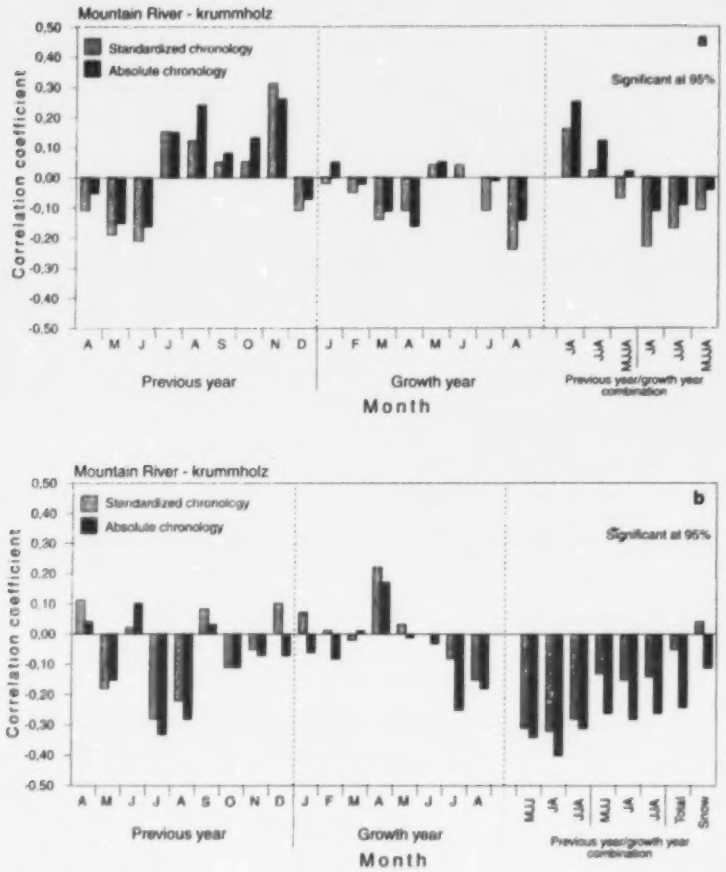


Figure 13.

Correlation of tree growth with ***a)*** *monthly mean temperature and* ***b)*** *total monthly precipitation at Norman Wells (1944–1989) for the standardized and absolute Mountain River krummholz chronology.*

averaging monthly temperatures from groups of months provides a stronger negative correlation between summer temperatures preceding ring formation (May to August) and ring width ($r = -0.49$). This relationship is apparent in Figure 14a, in which average summer temperatures prior to growth and the inverse of the tree-ring indices are plotted on same scale ordinates. Using the inverse indices when the correlation is negative makes the correspondence more visible. On the other hand, correlations between forest-site tree growth and precipitation show no clear trends. The only determinant variable is precipitation in the preceding May, which has a negative effect on growth. Precipitation during the previous summer (June to September) and during the current spring (April), although not significant, is positively correlated with ring width. The opposite response of growth to temperature and precipitation seems typical for dry sites in subarctic environments (Villalba and Veblen, 1994). In fact, as noted by other workers (Bongaonkar et al., 1994), excessive solar radiation associated with higher temperatures during summer has a direct effect on evaporation and the depletion of soil moisture. Since higher temperatures are physiologically more conducive to better growth in boreal environments, the negative response here is probably due to the dominant role of temperature in determining available moisture for the trees. Finally, the analysis of the climate-growth relationship also suggests that warm and dry springs (especially in April) could be critical for tree growth, probably by increasing desiccation damage (D'Arrigo et al., 1992). For those reasons, fluctuations in the Mountain River forest chronology should be interpreted more in terms of water stress than strictly in terms of summer temperatures. The increasing radial growth at the end of the nineteenth century reflects increasing precipitation and soil moisture associated with the twentieth-century warming (Jones et al., 1986). The suppressed growth conditions which characterize the Little Ice Age (Grove, 1988), and particularly the period of 1780–1850, indicate a strong water stress in trees.

The situation is quite different at the krummholz site where tree growth exhibits a negative correlation with previous and current summer precipitations and positive correlation with previous summer (July to August) and fall (November) temperatures (Fig. 13). The significant negative correlation between summer precipitation and radial growth is also obvious in Figure 14b where total July–August precipitation of the previous year is plotted against the inverse of the tree-ring indices. Climatic conditions during the year prior to growth seem significant at this exposed site. The negative response of radial growth to summer precipitation indicates that such stunted trees, characterized by a reduced leaf surface, are not sensitive to water stress during the summer. The krummholz response to climatic factors is thus similar to the

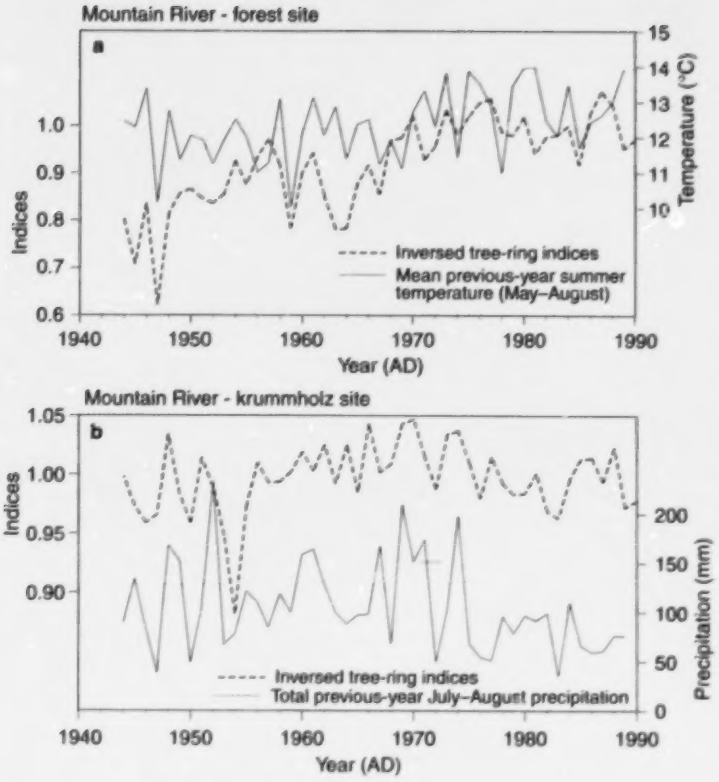


Figure 14.

a) *Comparison between absolute tree-ring indices (inversed) of forest chronology and mean summer temperature (May–August) of the previous year.* ***b)*** *Comparison between absolute tree-ring indices (inversed) of krummholz chronology and total July–August precipitations of the previous year.*

one observed at several treeline sites where tree growth is favoured by warm conditions during the previous summer. Note that moisture stress during the spring can represent a problem for further tree growth. The correlations of April temperature (negative) and precipitation (positive) with tree growth indicate that the initiation of photosynthesis when the soil is still frozen can increase desiccation processes and foliage damage. The physiological interpretation of the climate-growth relationship shows that variations in the tree-ring pattern at the krummholz site provides information about summer temperature since AD 1500. The krummholz chronology (Fig. 11) suggests that warm, and probably wet, conditions prevailed during the sixteenth century, before onset of the cooler period of the Little Ice Age (approximately 1600 to 1900). Most trees in the krummholz germinated during this mild period and developed a normal shape (single erect stem). The absence of erosion marks on subfossil trees suggests that the protective snow level was probably higher at this time. Conversely, the following lengthy depressed period on the tree-ring curve shows the 'Little Ice Age' cooling effect on tree growth. Summers were cooler and probably drier, while the protective snow cover had decreased substantially. In the same area, low snow level and the ensuing low water level prior to the twentieth century has allowed the stabilization of glaciofluvial slopes and, consequently, the cliff-top dune surface (Bégin et al., 1995). The increasing desiccation and winter abrasion seriously deteriorated the leaf structure of trees and would have been responsible for the krummholz development during these three centuries. Fortunately, trees were able to recover partially during short milder episodes centring around the mid-1600s and the early 1800s.

Since the early 1900s, warmer summers and increasing snow associated with the 'Twentieth Century Warming' and deduced from slope activity and dune dynamics (Bégin et al., 1995) allowed trees to increase their radial growth. As mentioned before, snow level in such an exposed environment is a key factor for tree growth. It determines the proportion of trees exposed to winter sand and ice abrasion and hence the overall shape of trees. When snow level increases, the number of terminal buds protected by snow also increases, allowing the development of new leaders. Because of the physiological relationship between the overall photosynthetic structure and radial growth, ring width significantly increases; however, this increase in growth indices was not as spectacular as in the 1500s, probably because most of trees were more than 400 years old and had a substantially reduced vegetative structure when the climate changed. Nevertheless, the positive reaction of such dying trees clearly indicates that environmental conditions, and particularly summer temperature, have changed substantially since 1920.

SYNTHESIS AND CONCLUSION

The results presented in this paper clearly show the potential of tree-ring series to provide high resolution (annual) proxy climate data that span the last few centuries or the last

millennium in the Mackenzie valley. Treeline environments in the Eskimo Lakes area are particularly promising for long-term reconstructions because of the presence of old-growth conifer stands and their sensitivity to environmental changes. Even if chronologies cover different time intervals and represent past conditions in contrasting environments separated by hundreds of kilometres, they show similarities in long-term climatic trends. Climatic trends over the last millennia in North America and Europe from several proxy climate indicators seem to have been recorded in tree-ring sequences in the Mackenzie valley. Suppressed growth between the mid-1500s and early 1900s and the following increasing growth is a feature of all three locations presented here (Fig. 15). The depressed growth period that is apparent since ca. AD 1330 in the Eskimo Lakes chronology, corresponds to the 'Little Ice Age' (Grove, 1988). This well known cold period has been responsible for significant glacier advances in the Canadian Rockies (Luckman, 1986, 1994) and for altitudinal tree-line regression (Kearney and Luckman, 1987). In marginal forest environments (e.g. Eskimo Lakes and Mountain River krummholz) the cooling effect seemed severe enough to transform forest sites into krummholz, mostly by decreasing snow level. In the Mountain River area, the tree growth-climate relationship indicates that, during the Little Ice Age, ring width has decreased in response to a moisture stress.

Depending on the site, some milder intervals occurred during the Little Ice Age. The period from approximately AD 1500–1550 has been the most important one and is recorded in both exposed sites (Eskimo Lakes and Mountain River krummholz). Warmer conditions centred in the sixteenth century have been recorded in other sensitive tree-ring sequences in the Rockies (Luckman, 1994) and in northern Quebec (Payette et al., 1989). The Mountain River chronologies do not reveal what has happened in the boreal forest before about 1500; however, the tree-line chronology suggests that the first half of the sixteenth century has to be viewed only as a brief warm interval within the cold Little Ice Age that seems to have prevailed since 1330. The Eskimo Lakes chronology also shows that the period from 1172 to 1330 was particularly favourable for tree growth. Overall, the highest tree growth values were recorded during this period, suggesting that summers were probably warmer than today. Moreover, the normal shape of subfossil trees associated with this period indicates that winters probably were also milder with a greater snowfall. Such environmental conditions probably correspond to the end of the period known as the 'Medieval Warm Period', which spans from about the ninth to the fourteenth centuries (Hughes and Diaz, 1994). The tree-ring signal from Eskimo Lakes seems consistent with historical records from western Europe which indicate the predominance of warm springs and dry summers throughout most of the thirteenth century (from AD 1220–1310). Finally, all tree-ring chronologies show a relatively abrupt increasing

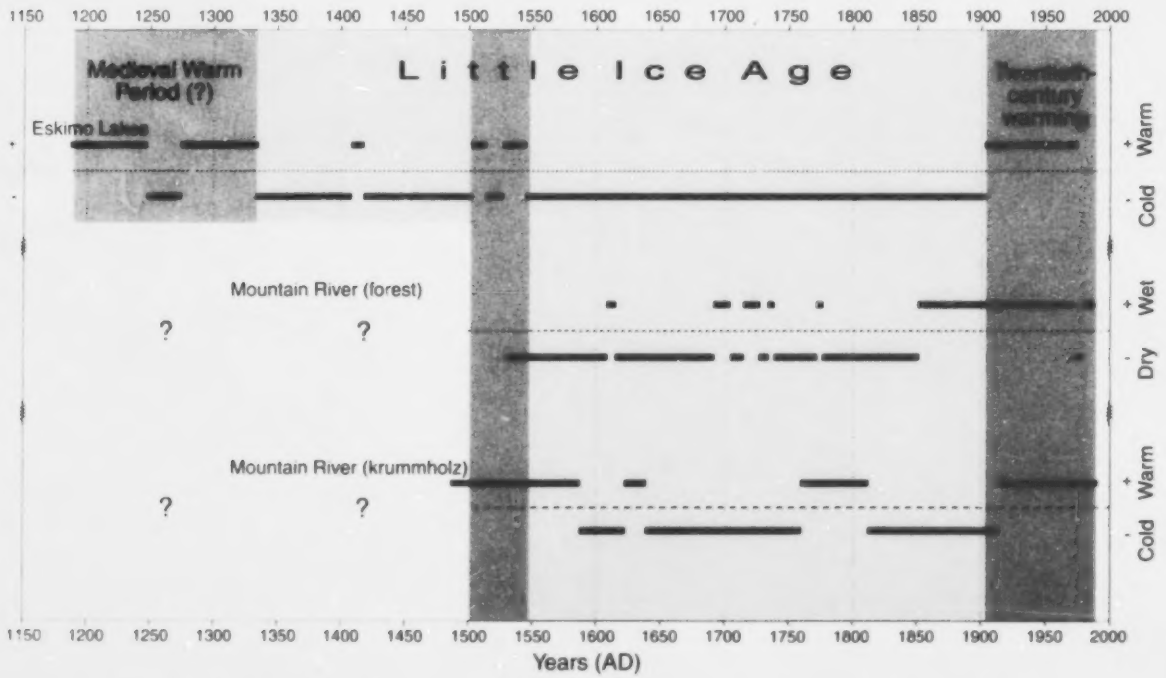


Figure 15. *Synthesis of the reconstructed climate changes based on tree growth–climate relationship for Eskimo Lakes and Mountain River sites. Tree-growth excursions above and below the mean are indicated by straight lines, while shades of grey emphasize broad-scale climate changes. Boundaries of broad climatic periods are based on tree-ring patterns presented in this paper.*

growth at the beginning of the twentieth century. In the forest site at Mountain River, the increasing growth started in mid-1800s while it started in early 1900s at the other two sites which are more exposed. The long-term warming trend since approximately 1880 is one of the most ubiquitous features in northern North American tree-ring records (Jacoby and Cook, 1981; Jacoby et al., 1985; Payette et al., 1985; D'Arrigo and Jacoby, 1992; Szeicz and MacDonald, 1994). It has also been recorded in cores from the southern Greenland ice cap based on predominance of melt layers (Bradley and Jones, 1993), and in recession rates of alpine glaciers in Alaska (Heusser and Marcus, 1964).

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Past environmental change recorded in dune fields

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Michaud, Y. and Bégin, C., 2000: Past environmental change recorded in dune fields; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 79–87.*

Abstract : Sand dunes are excellent markers of past eolian activity and paleoenvironmental conditions. In the Mackenzie valley, sand dunes mainly occur in the vicinity of Fort Simpson, in the region of Tulita to Fort Good Hope, and on the northern tip of Tuktoyaktuk Peninsula. The occurrence of both active and stabilized dunes illustrates that paleoenvironmental conditions controlling their development must have changed at least locally during the Holocene. Two generations of dune fields are recognized: old dune fields that were formed during the Late Wisconsinian, and younger dune fields that have been episodically or continuously active in the Holocene. Comparison between dunes morphology, composition, orientation, and age with wind-direction and temperature-moisture records revealed that caution must be exercised in placing a climatic interpretation upon eolian phenomena at regional scale. In many locations, eolian activity may be triggered by ecological and geomorphic factors as well as climatic change.

Résumé : Les dunes de sable sont d'excellents indicateurs de paléoérosion éolienne et de conditions paléoenvironnementales. Dans la vallée du Mackenzie, des dunes de sable se sont formées près de Fort Simpson, dans la région située entre Fort Norman et Fort Good Hope, ainsi qu'à l'extrémité nord de la péninsule de Tuktoyaktuk. La présence de dunes actives et de dunes fixes indique que les conditions environnementales qui prévalaient pendant leur évolution se sont modifiées, du moins localement, durant l'Holocène. On distingue deux générations de champs de dunes : les champs de dunes formés durant le Wisconsinien supérieur et les champs de dunes qui étaient actifs de façon épisodique ou continue durant l'Holocène. La comparaison entre la morphologie, la composition, l'orientation et l'âge des dunes, d'une part, et la direction du vent, la température et l'humidité, d'autre part, incite à dedoubler de prudence lorsqu'on interprète les phénomènes éoliens à l'échelle régionale. En nombre d'endroits, l'activité éolienne peut être due à des facteurs écologiques et géomorphologiques ainsi qu'à un changement climatique.

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INTRODUCTION

Sand dunes are excellent markers of past eolian activity and paleoenvironmental conditions. Dunes may reveal information on vegetation, fire history, temperature-moisture regimes, and wind regimes (David, 1981; Filion, 1984; Thorson and Bender, 1985); however, caution must be exercised in placing a climatic interpretation upon eolian phenomena, in recognition that the direct controls may be ecological or geomorphic factors that may or may not be related to climate. Lake drainage, river-level fluctuations and slope movements are processes that may contribute to sand-dune development by exposing sandy material to wind erosion.

Dune fields occur throughout Canada (David, 1977, 1981), usually in the vicinity of glaciofluvial deltas and outwash plains and glaciolacustrine deposits. Although dune fields have been documented in both subarctic and arctic regions (e.g. Carter, 1981; Filion et al., 1991), very little information has been published about sand dunes in the Mackenzie valley. David (1977) briefly described the dune fields of the Fort Simpson area, while Mackay (1963), Rampton (1988), and Ruz (1993) discussed the occurrence of eolian sediments and foredunes in the Tuktoyaktuk area. This paper provides an overview on the nature, extent, and age of major dune fields in the Mackenzie valley and a brief discussion of the paleoenvironmental changes recorded in dune-field history.

DISTRIBUTION AND AGE OF MAJOR DUNE FIELDS

Dune fields are rare in the Mackenzie valley. They mainly occur in the vicinity of Fort Simpson, and in the region of Tulita (formerly Fort Norman) to Fort Good Hope. A major dune field also lies along the Beaufort Sea coast on the northern Tuktoyaktuk Peninsula (Fig. 1). In the Mackenzie valley, dune fields are mainly composed of parabolic and cliff-top dunes. Most of the dune fields are stabilized with a forest cover, but portions of several dune fields are presently active. The occurrence of both active and stabilized dunes illustrates that conditions controlling their development must have changed at least locally during the Holocene.

Fort Simpson dune fields

Located in the southern Mackenzie valley, the Fort Simpson dune fields consist of three adjacent fields that cover 700 km^2 on both sides of the Mackenzie River near the confluence of Liard and Trail rivers (Fig. 1). They are mainly composed of fossil parabolic dunes and irregular hummocks.

Sand dunes several hundred metres in length and 10–15 m high are widely distributed in this area (Craig, 1965; David, 1977). Field observations along roadcuts on a raised delta of the ancestral Liard River reveal that dunes are composed of fine- to very fine-grained sand and do not contain basal or intercalated organic layers. Figure 1a shows the majority of the dunes that are large enough to be seen on airphotos (J. Aylsworth, pers. comm., 1993). A large number of the dune crests are oriented southeast-northwest, whereas some of them are arcuate towards the west. Other sand ridges oriented southwest-northeast are also present in the area, but considering that they are at least twice as large as the eolian dunes and that many of the ridges show small tails almost perpendicular to their orientation, they are interpreted as alluvial bars associated with the Liard River deltaic formation that have been reworked by eolian activity.

Detailed observation of the southeast-northwest dunes reveals that all of them are asymmetrical parabolic dunes where many show signs of a northwesterly influencing wind and some show evidence of a southeasterly influencing wind. Based on the morphology of all eolian features occurring in the Fort Simpson dune fields, dune-constructing winds were likely coming from the northwest, southeast, and east. Comparison of dune-crest orientations with the wind rose diagram from the Fort Simpson weather station for the interval 1964 to 1990 (Fig. 1) shows that both the southeasterly and the northwesterly winds are still present, whereas easterly winds are rare.

None of the dunes in the Fort Simpson area are presently active. They are heavily forested with the interdune areas being covered by swamps and bogs. Because little buried organic matter occurs within the sand hills, thermoluminescence dating was used to obtain ages on sand grains sampled from easterly and northwesterly oriented dunes. The method known as optically stimulated luminescence (OSL; *see* Aitken and Xie, 1992) was used to determine an age of 8200 ± 1000 years (*see* Appendix 2 for information on this date) for the surface (30 cm depth) of a parabolic dune chain situated on the northwest side of the Liard River and an age of 9000 ± 1000 years (*see* Appendix 2) for sand grains located at 5 m depth in a 15 m high sand ridge situated near Manners Creek.

The morphology, composition, and age of the sand hills indicate that the dune fields were produced in a single phase of eolian activity. The optically stimulated luminescence dates suggest that this eolian period must have started after the drainage of glacial Lake Mackenzie, right after deglaciation (11 500 BP) (Smith, 1992; Lemmen et al., 1994), until approximately 8000 years BP. Drainage of glacial Lake Mackenzie would have exposed the surface of the ancestral Liard River delta to wind erosion eventually leading to the formation of the dune fields. The eolian activity lasted a few thousand years, after which the dunes became stabilized by vegetation. There is no evidence of reworking or reactivation of eolian activity in the Fort Simpson area after this early period of dune formation. This extensive period of eolian activity was characterized by southeasterly, northwesterly, and easterly dune-building winds generated by an anticyclonic air mass centred on both the Laurentide Ice Sheet (David, 1981) and the Cordilleran Ice Sheet, and by an abundance of sand and silt deposits that were exposed as a result of both the retreat of the Laurentide Ice Sheet and the demise of glacial Lake Mackenzie.

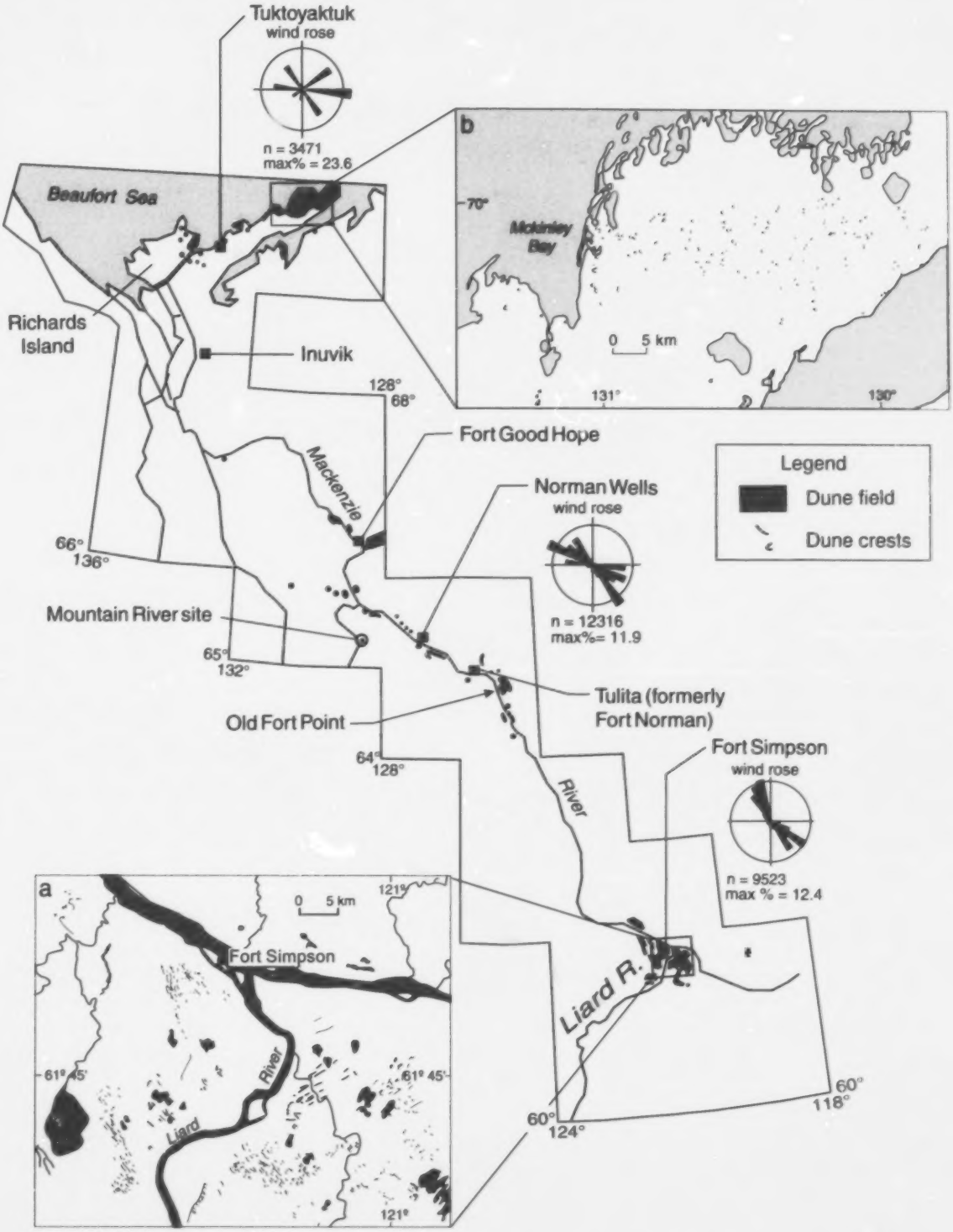

Figure 1. *Location and distribution of dune fields in the Mackenzie valley with enlarged maps of* ***a****) dune distribution in the Fort Simpson area, and* ***b****) dune distribution on Tuktoyaktuk Peninsula. Symbols representing parabolic inland dunes on Tuktoyaktuk Peninsula are arcuate in the downwind direction. Bars on the wind rose diagrams indicate the dominance of wind direction in 10° increments.*

Central Mackenzie valley dune fields

Numerous small dune fields or isolated dunes are also present in the central portion of the Mackenzie valley. Most are concentrated along the Mackenzie River between Old Fort Point, near Tulita, and Fort Good Hope, while a few others are distributed sporadically along tributaries. All the dune fields near the Mackenzie River are found on glaciofluvial and glaciolacustrine sediments in the vicinity of glacial Lake Mackenzie (Duk-Rodkin and Hughes, 1992, 1993; Smith, 1992). No dates are available for dune activity at these sites, but considering their location they probably formed immediately after the lake drainage, concomitant with the Fort Simpson dune fields. Individual fields, and even individual dunes, were formed at slightly different times related to the gradual regression of glacial Lake Mackenzie, exposing the former lake bed.

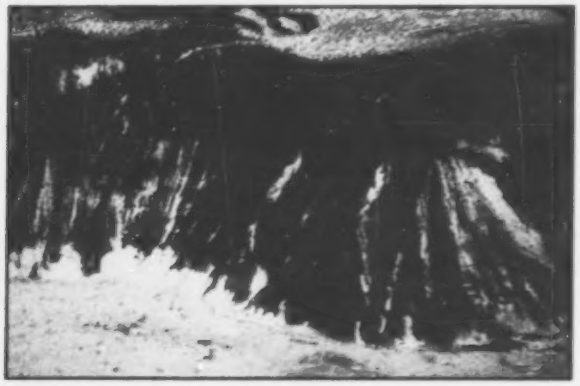

Figure 2. *Oblique view of Mountain River cliff-top dune. The winter photograph illustrates the niveo-eolian character of that dune. It also illustrates that the dune is well developed only above a snow- and vegetation-free escarpment. The cliff is 85 m high and the dune is approximately 160 m wide. Photograph by M. Nixon. GSC 2000-029*

In the central Mackenzie valley, a cliff-top dune along Mountain River provides an opportunity for paleoenvironmental reconstruction for this area over the last millennia (Bégin et al., 1995). The expansion of this dune towards the north-northeast into a mature white spruce stand along with buried organic horizons and dead trees in growth position, permits an assessment of cliff-top dune dynamics (Fig. 2).

The cliff-top dune, developed on the western edge of a glaciofluvial terrace, covers approximately 11 300 m^2 and consists of a bowl-shaped deflation zone exposing four organic layers, a 13 m high dune crest, and a 3200 m^2 accumulation zone strewn with dead trees buried in vertical position (Fig. 3). Initial cliff-top dune formation occurred between AD 750–860 when sand sprinkling improved the radial growth of trees located within D1-D2 organic horizons. Addition of sand through sand sprinkling modified the D1-D2 forest soils enough to allow trees to produce larger tree rings (Fig. 4). Dune building was actually initiated

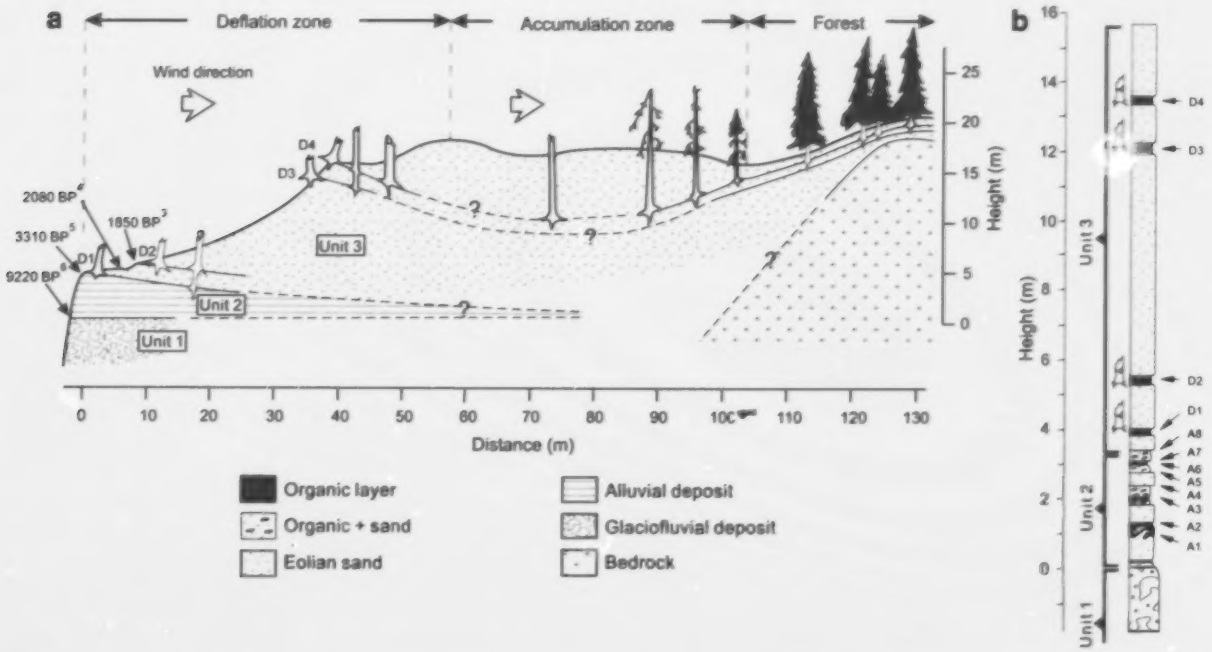

Figure 3. ***a)*** *Schematic longitudinal section of the Mountain river study site showing the stratigraphic units, the position of the forest paleosols (D1 to D4) and the vegetation types of the cliff-top dune.* ***b)*** *Stratigraphic section; A1–A8 refer to organic layers within the alluvial unit.* See *Appendix 2 for list of laboratory numbers for dates given in this figure.*

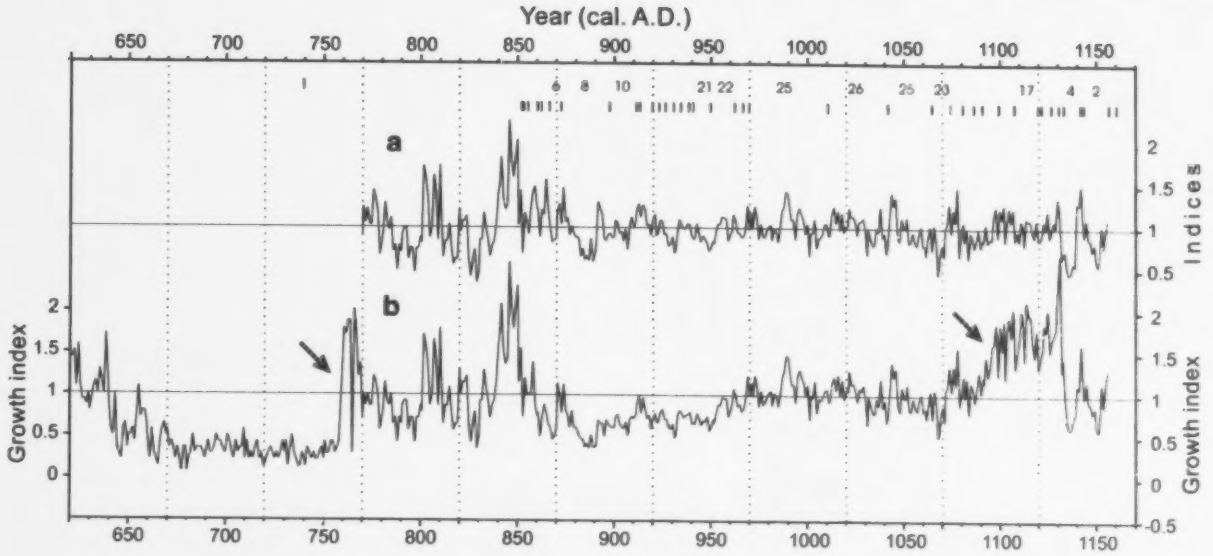

Figure 4. *a) Standardized, and b) growth-indexed series from dead trees (n = 26) from D1 and D2 forest paleosols at the Mountain River site. Arrows along the growth-index curve (b) are pointing at periods where tree-ring widths were getting larger than usual in response to slow eolian sedimentation. Numbers at the top refer to the total number of wood samples available. Calendar dates were derived from calibrated radiocarbon dates obtained from subfossil trees.*

around AD 1100, causing the death of several D1-D2 trees and eventually the burial of the forest soils under 7 m of sand. This eolian period lasted about 300 years and was followed by a stabilization phase which is recorded by the arrival of D3-D4 trees around AD 1460. This surface stabilization lasted a few hundred years until AD 1860–1865, when a second episode of eolian activity began.

An average sedimentation rate of 34 cm/year and an accumulation-front displacement of 78 cm/year for the dune were obtained from direct measurements during the summers of 1990 through 1992. Conversely, average erosion rates of 65 cm/year and 60 cm/year were calculated respectively for micro-cliff retreat and surface lowering of the windward zone. Finally, the presence of snow layers interbedded with sand layers and evidence of snow removal in the accumulation zone suggest that winter sand transportation is very important at the site. It appears that much of the dune construction occurs in winter and is the result of niveo-eolian activity.

Tuktoyaktuk area dune fields

The northern tip of the Tuktoyaktuk Peninsula is largely covered by a veneer of eolian sediments (Mackay, 1963; Rampton, 1988). Dune fields run across the peninsula in the vicinity of McKinley Bay and also occur on the eastern shore of Richards Island. Isolated sand dunes occur in the Tuktoyaktuk region between the eastern tip of the peninsula and the eastern end of Richards Island (Fig. 1; Mackay, 1963; Rampton, 1988). The largest dune field lies on the northern tip of the Tuktoyaktuk Peninsula and covers approximately 500 km^2 (Fig. 1b).

The Tuktoyaktuk Peninsula dune field is characterized by the presence of three types of dune that developed in different geomorphic settings and at different times. First, there are the 'inland dunes' which are low-lying parabolic dunes with average lengths varying from 300 m to 600 m and heights ranging from 0.5 m to 3 m. Most of these dunes are arcuate towards the west, thus indicating that they were constructed by easterly winds. These dunes are composed of fine-grained sand (average sand-grain diameter = 0.15 mm (2.85 ϕ)) and developed upon a sandy outwash plain (Rampton, 1988). They are fossilized features that are stabilized by a 25 cm thick humus layer topped by tundra vegetation cover (Fig. 5). An optical date (optically stimulated luminescence) for a fossil parabolic dune produced an age of 8600 ± 1000 years (*see* Appendix 2). This is one of the oldest dates for eolian activity in the Tuktoyaktuk area and is believed to represent the end of dune formation because the dated sand was sampled near the surface.

Second, there are numerous 'lakeshore dunes' located around drained thermokarst lakes. Some are blowout dunes and others are simply elongate ridges developed atop a low scarp surrounding lake edges (Ruz, 1993). Many drained thermokarst lakes have their own group of dunes formed from sand derived from the emerged lake shelves. In general, dunes are located on the eastern, southern, and western sides of partially to completely drained lakes. The development of these dunes depends on the catastrophic drainage of the individual thermokarst lakes, which occurs on the Tuktoyaktuk Peninsula through localized melting of permafrost and coastal retreat due to sea-level rise (Mackay, 1986, 1988; Ruz, 1993). The ages and periods of activity of the lakeshore

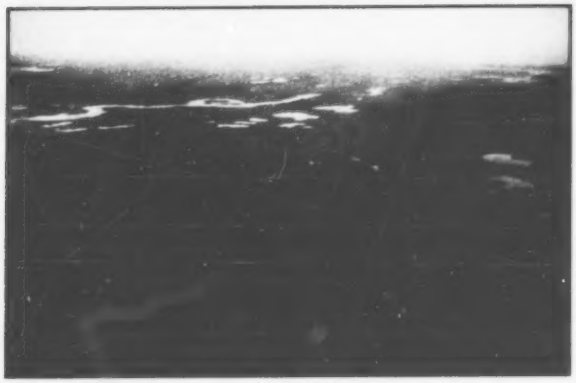

Figure 5. Fossil parabolic dune on Tuktoyaktuk Peninsula. This dune is typical of many inland dunes observed near McKinley Bay. The foreground dune is in the order of 400 m in length and 0.5 m in height. Photograph by Y. Michaud. GSC 2000-030

dunes are variable and dependent upon individual lake drainage, which controls the availability of sand. Detailed dating of organic layers intercalated in lakeshore dune sequences in Atkinson Point (Ruz, 1993) and McKinley Bay (Bégin and Michaud, 1993) suggests periodic activity of these dunes beginning around 3000 years BP (Fig. 6). These organic layers represent several phases of dune stabilization, indicating that these dunes evolved separately in response to drainage of their nearby thermokarst lakes.

Third, there are 'coastal dunes' bordering the Beaufort Sea. These dunes occupy a narrow coastal strip and are widely distributed along Tuktoyaktuk Peninsula and Richards Island (Rampton, 1988; Ruz, 1993). Foredunes are found landward of large beaches along low shorelines, while cliff-top dunes are formed above some of the retreating sand bluffs. These are subjected to both easterly and strong westerly winds usually associated with storms. At present, all coastal dunes are active. Given the context of shoreline erosion caused by the Holocene sea-level rise (Hill et al., 1993), it may be argued that coastal dunes likely formed continuously throughout the Holocene.

PALEOENVIRONMENTAL RECONSTRUCTION ON THE BASIS OF EOLIAN ACTIVITY

Two generations of dune fields are recognized in the Mackenzie valley: 1) old dune fields that were formed during the Late Wisconsinan (prior to 8000 BP), and 2) younger dune fields that have been episodically or continuously active in the Holocene.

Numerous dune fields were developed at the periphery of the Laurentide Ice Sheet shortly after its retreat (Gadd et al., 1971; David, 1981; Filion, 1987; Côté et al., 1990), suggesting the presence of katabatic winds emanating from an anticyclonic air mass centred on the ice sheet (David, 1981). This

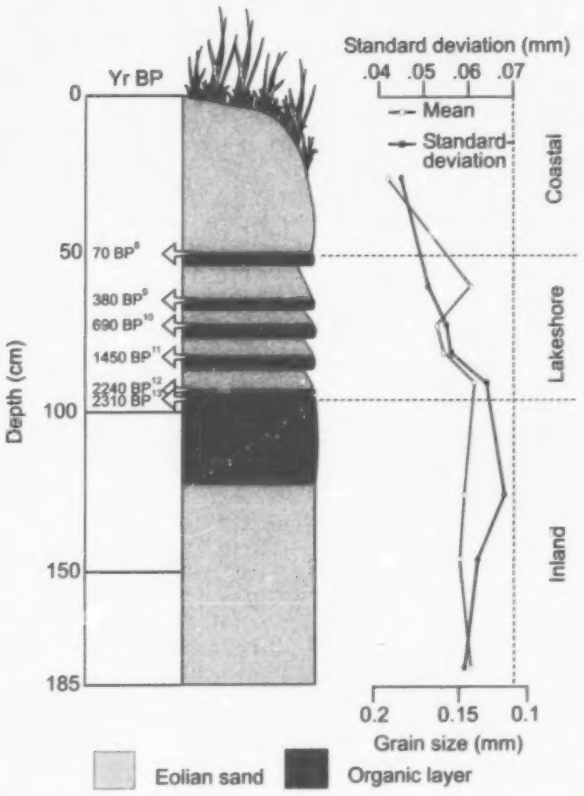

Figure 6. Cross-section showing the detailed chronostratigraphy of a sand bluff along McKinley Bay shoreline. Notice the upward evolution of dune type as coastal retreat progresses towards this location. See *Appendix 2 for list of laboratory numbers for dates given in this figure.*

wind regime seem to have lasted about 2000–2500 years before the temperature-moisture regime of these areas was altered following ice-sheet breakup (David, 1981; Filion, 1987). As shown in Figure 7, the orientation of the old dune fields of the Mackenzie valley indicates that they were formed by bidirectional winds in a large corridor confined between two ice caps. Katabatic winds induced by both the Laurentide and the Cordilleran ice sheets seem to be responsible for the formation of these dune fields during the Late Wisconsinan. Since paleoenvironmental reconstructions for the Mackenzie valley (Spear, 1983; Ritchie, 1984, 1985) revealed that the temperature-moisture regime of the valley had become gradually warmer and more humid by 12 000 BP, they indicate that Fort Simpson and some central Mackenzie valley dune fields likely followed the same pattern of development as the dune fields described in the Prairies and in St. Lawrence Lowlands. As for the Tuktoyaktuk Peninsula inland dunes, the optically stimulated luminescence date suggests that these dunes began forming after the retreat of the Laurentide Ice Sheet, but the orientations of the dunes (Fig. 1b) show that they were built by easterly winds. Therefore, these inland dunes formed under a polar easterly wind regime different from that operating in the Mackenzie valley.

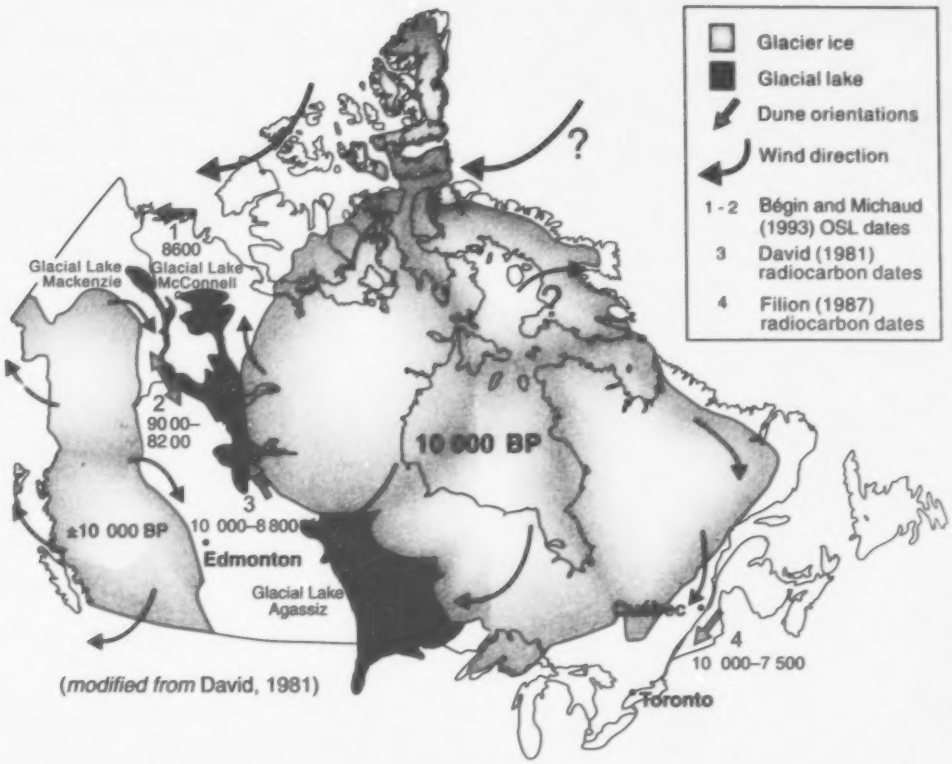

Figure 7. *Position of some dune fields that were active during the Early Holocene. Most of them were located adjacent to the Laurentide Ice Sheet and were under the influence of anticyclonic air-mass circulation centred over the Laurentide and the Cordilleran ice sheets. Notice that late Wisconsinan inland dunes on the Tuktoyaktuk Peninsula were formed under a polar easterly wind system, which was independent of the wind pattern related to the ice sheets.*

Thus, the late Wisconsinan period seems to have been marked by the presence of two dominant wind systems, namely the anticyclonic air-mass systems centred on both the Laurentide and the Cordilleran ice sheets and the polar easterlies (Fig. 7).

According to wind-direction records from Fort Simpson, Norman Wells, and Tuktoyaktuk (Fig. 1), air masses centred on the ice sheets have been more or less replaced by a wind regime that follows the overall topography of the valley, while the polar easterlies remain active in the Mackenzie Delta. In fact, comparisons between contemporary wind patterns and Holocene eolian activity in the Mackenzie valley, as well as on the Alaskan arctic coastal plain (Carter and Robinson, 1978; Carter, 1981, 1993), indicate that arctic wind regimes have not changed much during the Holocene.

The activity of younger dunes is more difficult to assess regionally, because they are often represented by systems that are active either continuously or episodically, depending upon local factors which control the availability of sand. This seems to be the case for the Mountain River cliff-top dune, where downcutting and slope processes were key triggering mechanisms, and also the case for lakeshore and coastal dunes on the Tuktoyaktuk Peninsula where thermokarst lake drainage and shoreline erosion control the activity; however, by compiling radiocarbon dates for dune initiation in locations in the Arctic, it seems that eolian activity resumed around 4000 years ago resulting from the deterioration of climate in the Arctic (Ritchie and Hare, 1971; Pissart et al., 1977). Dune inception dates of 3790 ± 90 BP and 3460 ± 80 BP from Banks Island (Pissart et al., 1977), 3280 ± 130 BP (Lowdon and Blake, 1978), 2310 ± 70 BP (*see* Appendix 2) (Bégin and Michaud, 1993), and 1530 ± 70 BP (Ruz, 1993) from the Tuktoyaktuk Peninsula, plus many others from northern Alaska (Galloway and Carter, 1992) have been used to interpret the resumption of eolian activity around the Beaufort Sea. On the other hand, the renewal of eolian activity around 4000 BP does not mean that sand dunes were absent in the region prior to that date. Ages of 8430 ± 120 BP (Pissart et al., 1977) and 5800 ± 180 BP (Pissart et al., 1977) were reported from Banks Island, while coastal dunes would likely been active throughout the Holocene.

Obtaining significant regional paleoenvironmental records from younger and smaller dune systems is difficult. It has been shown that many of these dunes are subject to local perturbations controlled by sediment supply. Lakeshore dunes on the Tuktoyaktuk Peninsula and cliff-top dunes along river banks are good examples of such situations. The chronology of lakeshore dunes seems to highlight thermokarst-lake water-level fluctuations (Ruz, 1993) whereas the chronology of the cliff-top dune on Mountain River provides indirect information on river-level fluctuations, slope processes, and katabatic wind regime. Before significant climatic signals can be interpreted from these dune systems, periods and triggers of eolian activity need to be validated by the chronology of other geomorphic processes from the entire Mackenzie valley.

ACKNOWLEDGMENTS

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Appendix 1

Temperate sand-dune nomenclature

A sand dune can be defined simply as a hill or a ridge of sand piled up by the wind (Pye and Tsoar, 1990); it can form under any climatic regime, the key conditions being that bare sand is exposed to wind that is strong enough to entrain sand particles. Deposition of sand at any location is then governed by a combination of factors such as topography, bed-roughness changes, or aerodynamic fluctuations and vegetation. In general, eolian processes are more important in deserts, but they also occur in temperate regions where sand moisture can vary through time and from place to place. In Canada, most, if not all, dunes formed under temperate conditions.

The formation of sand dunes requires the occurrence of three basic elements: 1) a sand-transporting wind (windspeed > 5 m/s), 2) some bare sands available for wind transportation, and 3) some sort of obstacles to trap the sand. Since these requirements are often met, sand dunes are found in many geographic locations, climatic zones, and geographical settings. Sand dunes have been classified and named on the basis of criteria related mainly to morphology and geographical setting.

Morphological aspects

In temperate zones, all dunes are basically **parabolic dunes**. They are curved dunes where sand is removed from the centre of the dune (blowout zone) and carried downwind, leaving two arms pointing upwind. This type of dune is also known as a blowout dune. Depending on the amount of sand available, parabolic dunes may become highly deformed, generating several forms of dunes such as **asymmetrical**, **elongated**, **fish-hook**, **V-shaped**, and **dune ridges**. These are a few of the forms which result from a lack of sand supply.

Geographical settings

Sand dunes are often classified in accordance with the environment or surrounding landscape where they develop. Dunes such as **coastal dunes**, **lake-shore dunes**, and **inland dunes** refer to the general environment where they formed, whereas dunes such as **cliff-top dunes** and **foredunes** refer to a specific topographic context.

Appendix 2

List of laboratory numbers for dates given in text

Number	Age	Standard deviation	Laboratory
1	8200 ± 1000 BP	1	LUX-104gf
2	9000 ± 1000 BP	1	LUX-103gf
3	1850 ± 100 BP	2	GSC-5348 *
4	2080 ± 70 BP	2	GSC-5358 *
5	3310 ± 90 BP	1	BETA-50209
6	9220 ± 170 BP	1	BETA-50210
7	8600 ± 1000 BP	1	LUX-105gf
8	70 ± 80 BP	2	GSC-5573 *
9	380 ± 70 BP	2	GSC-5563 *
10	690 ± 60 BP	2	GSC-5551 *
11	1450 ± 70 BP	2	GSC-5545 *
12	2240 ± 80 BP	2	GSC-5537 *
13	2310 ± 70 BP	2	GSC-5525 *

* $\delta^{13}C_{PDB} = -25‰$

Shallow ground temperatures

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Burgess, M.M. and Smith, S.L., 2000: Shallow ground temperatures; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 89-103.*

Abstract: The temperature in the ground changes according to daily or longer temperature cycles in the air. The amount of ground temperature change also diminishes rapidly with depth. Temperatures to a depth of 20 m, the approximate depth to which the yearly temperature cycle penetrates, are presented for various environments in the Mackenzie valley. As a rough guideline, average ground temperatures are about 4°C warmer than mean annual air temperatures. This difference depends mainly on the insulating affect of vegetation and snow and changes in the moisture content of the active layer. Smaller differences may result in peatlands where summer drying of organic soils enhances their insulating capacity. Even without a change in climate, disturbance at the ground surface will alter the ground thermal regime. Examples from the Norman Wells to Zama, Alberta oil pipeline show warming and ground subsidence associated with the clearing of the pipeline right-of-way.

Résumé : La température du sol dépend des cycles de la température de l'air, quotidiens ou à plus long terme. L'amplitude de la variation de la température du sol diminue rapidement avec la profondeur. Dans différents environnements de la vallée du Mackenzie, nous présentons les températures régnant à une profondeur de 20 m, la profondeur approximative à laquelle se font sentir les effets du cycle annuel de température. Grosso modo, les températures moyennes du sol sont plus élevées d'environ 4 °C que les températures annuelles moyennes de l'air. Cette différence est principalement attribuable à l'effet isolant de la végétation et de la neige et aux variations de la teneur en humidité du mollisol. On peut observer des écarts moins grands dans les tourbières où l'assèchement des sols organiques en été accroît leur capacité d'isolation. Même sans changement climatique, la perturbation de la surface du sol modifie le régime thermique du sol. À titre d'exemple, on observe un réchauffement du sol le long de l'oléoduc entre Norman Wells et Zama (Alberta), ainsi qu'une subsidence de la surface du sol causée par le dégagement du tracé.

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INTRODUCTION

An understanding of the current relationship between permafrost and climate is necessary before attempting to predict the impact of climate change on permafrost in the Mackenzie valley. The atmospheric climate is the main factor determining the existence of permafrost; however, the spatial distribution, thickness, and temperature of permafrost is highly dependent on the temperature at the ground surface. The temperature at the ground surface, although strongly related to climate, is influenced by several other environmental factors such as vegetation type and density, snow cover, drainage, and soil type. This paper characterizes ground temperatures within the depth range in which yearly changes take place, i.e. above 20 m.

Concepts and features of permafrost, and of the shallow ground thermal regime are first introduced. The broad relationship between climate and ground-surface temperatures in the Mackenzie valley is then outlined. Details on the great variability in the shallow ground thermal regime which exists due to differences in local environmental and site conditions are presented next, and illustrated with numerous examples from Geological Survey of Canada (GSC) study sites. Finally, some examples of the thermal response to surface disturbances are given. These examples may serve as an analogue for the effect of a sudden climate warming.

CHARACTERISTICS OF THE GROUND THERMAL REGIME

Temperature in the ground is directly related to the balance between incoming heat from the sun and the amount of heat re-radiated or reflected from the ground surface and consumed by evapo-transpiration. A small amount of heat also originates from the Earth's interior and, in conjunction with the thermal properties of subsurface materials, determines the maximum depth that permafrost can reach. Because this heat balance also controls air temperature, air temperature presents an accurate and easily measured quantity to relate to ground temperature. The relationship between the two is close, but not identical, because of the insulating effect of snow and vegetation.

Daily changes in air temperature penetrate only a few tens of centimetres below the ground surface. This tendency to dampen air-temperature change is a fundamental ground property. Snow, water, and vegetation between the ground and air add substantially to this damping. Only the annual change in air temperature from summer to winter is observed at depths of a few metres and even this is usually completely dampened out by a depth of approximately 10 m. This depth, where no annual cycle occurs, is known as the depth of zero annual amplitude. Below this depth, a systematic change in average annual temperature must take place, i.e. a change in climate, to cause a ground-temperature change. Systematic shifts in average annual ground temperature, whether observed within or below the depth of zero annual amplitude, can be used to infer past climate changes. An example of this is described in Dyke (2000a).

An example of the annual ground-temperature variation in a permafrost setting is shown in Figure 1. The maximum and minimum temperature experienced at each depth during the year define the annual ground-temperature envelope. The average temperatures at each depth define the mean annual ground-temperature profile. A year-long record for selected depths at another site is shown in Figure 2, along with the corresponding air-temperature record. Daily air-temperature

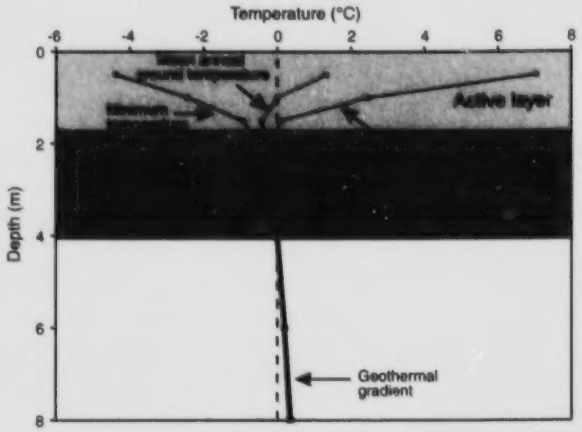


Figure 1. *Features of the ground thermal regime in permafrost, illustrated with data from a borehole in mineral soil near Fort Simpson, with thin permafrost. Note: The slope of the temperature line below the depth of zero annual amplitude is the geothermal gradient (assuming climate is stable at the ground surface).*

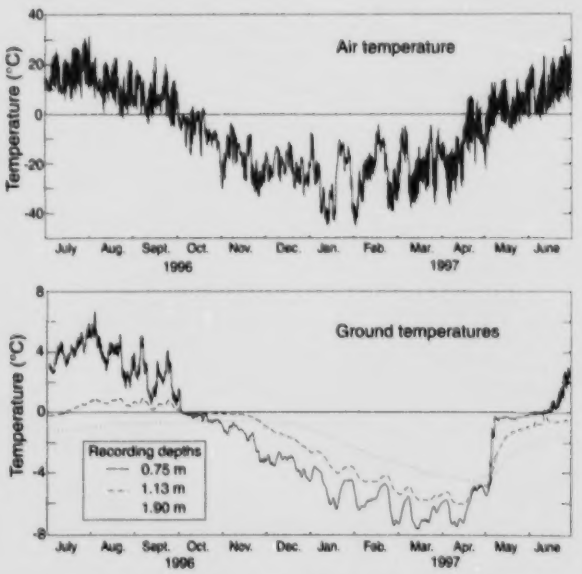


Figure 2. *Air temperature and near-surface ground temperatures recorded from June 1996 to June 1997 at a forested site on a slope, near Saline River, north of Wrigley.*

changes are completely removed by a depth of 1 m. As depth increases, the temperature changes become increasingly smooth, decreasing in amplitude and lagging progressively behind those of air temperature.

In permafrost environments, the layer of ground that freezes and thaws each year is known as the active layer, i.e. the zone where the maximum temperatures during the year exceed 0°C. (*see* Nixon (2000) for a discussion of active layers in the Mackenzie valley). The layer beneath the active layer that remains permanently below 0°C is permafrost. In permafrost environments, the mean annual temperature profile frequently exhibits a slight curvature towards higher temperatures within the active layer. This phenomena, known as the thermal offset, occurs due to differences in thermal properties of frozen and unfrozen ground. The upper part of the active layer may become particularly warm in summer, but this warmth will be restricted from penetrating deeply by the relatively low thermal conductivity of thawed ground. The thermal offset is also influenced by insulating snow cover (Goodrich, 1978, 1982; Burn and Smith, 1988). As a result of the offset, the mean annual ground temperatures at the surface and within the upper active layer are higher than those in the underlying permafrost (differences of over 1°C have been observed). In fact, the mean annual ground surface temperatures may be above 0°C, yet permafrost may exist. The temperature at the base of the active layer is thus more critical to the existence and dynamics of permafrost than the temperature at the ground surface (Smith and Riseborough, 1996).

Wet soils have quite different thermal properties than dry soils, and hence will respond differently to changes in surface temperature. A soil will have a higher heat capacity (the amount of heat required to raise the temperature of a unit volume by one degree) when wet than when dry, and thus the soil will cool or warm more slowly when wet. As well, there are differences between the thermal properties of frozen wet soils and those of unfrozen wet soils. Thermal conductivity (the quantity of heat that will flow through a unit area of a substance in unit time under a unit temperature gradient) is greater for wet soils when frozen than when unfrozen, since the thermal conductivity of ice is about four times that of water. In addition, in finer textured soils, the soil moisture will freeze or melt gradually over a range of temperatures below 0°C. Most soils thus have characteristic freezing curves (relationship between water content and temperature below 0°C), which depend on grain size, moisture content, pore-water salinity, and pressure. Unfrozen water contents may therefore be considerable in fine-grained soils at temperatures of 1 or 2° below 0°C. For example, a saturated silty clay with a porosity of 35% may contain 20% unfrozen water at a temperature of -0.5°C. The presence of unfrozen water will also lead to temperature-dependent mechanical properties, as well as to temperature-dependent thermal properties.

During freezing or thawing, ground temperatures are held at the freezing point as latent heat of fusion (the amount of heat required to melt the ice, or freeze the water) is absorbed or released. This phenomena is known as the zero curtain effect (*see* Fig. 2, where ground temperatures at a depth of 0.75 m drop rapidly below freezing in October, while those below at 1.13 m, where moisture contents are higher, remain at the freezing point for a considerably longer period of time). The gradual phase change is the result of an apparently increased heat capacity, an effect caused by the additional heat removal required to freeze the moisture in the soil.

The actual relationship between the climate above the ground and the temperature below is a function of site-specific conditions which determine the amount of heat available at the ground surface (surface energy balance). Incoming solar energy is the dominant energy source, while a small portion arrives from geothermal heat from the Earth's interior (this is shown by the geothermal gradient in Fig. 1). The geothermal heat flux also varies depending on the geological environment and history. Factors such as slope aspect, surface vegetation (or roughness), soil moisture, snow cover, presence or absence of an insulating organic layer, drainage, and elevation greatly influence the local surface heat balance, and hence the amount of heat available for warming the ground.

NEAR-SURFACE GROUND TEMPERATURE

Air-temperature data from Atmospheric Environment Service weather stations (Atmospheric Environment Service, 1982) and near-surface ground-temperature data have been used to produce a map of mean annual air and ground-surface temperatures for the Mackenzie valley region (Fig. 3a). The mean annual air temperature is based on the 30 year climate normals for 1951–1980. The near-surface ground temperatures from 176 sites have been extracted from a ground-temperature database compiled for northern Canada (updated from Young and Judge, unpub. report, 1985). The near-surface ground temperature was either directly determined from observations in the top metre of the ground (the actual depth varies from site to site according to temperature sensor depth), or by extrapolation of the mean annual temperature profile at depth below the level of zero annual amplitude to intercept the ground surface (when only deeper temperatures were available).

The mean annual air temperature ranges from slightly greater than -4°C in the southern part of the region to less than -10°C in the north. Mean annual near-surface ground temperature ranges from about 3°C in the southern part of the valley to about -10°C in the north. Mean annual air and near-surface ground temperature are roughly correlated (Fig. 3b), but considerable variability exists due to all the ground-surface factors listed above. How both the air and ground temperature vary with latitude in the Mackenzie valley is shown in Figure 3c, which shows that although both air and ground temperature generally decrease with increasing latitude, ground temperatures are usually higher than air temperature, usually by 2 to 4°C.

The magnitude of the difference between air and ground temperature and the local variation in ground temperature are dependent, as stated above, on several factors such as snow cover, surficial materials, vegetation cover, moisture conditions, and exposure to wind. The ground temperature, therefore, may show considerable variation over a small area even though air temperature may show little variation. In the area

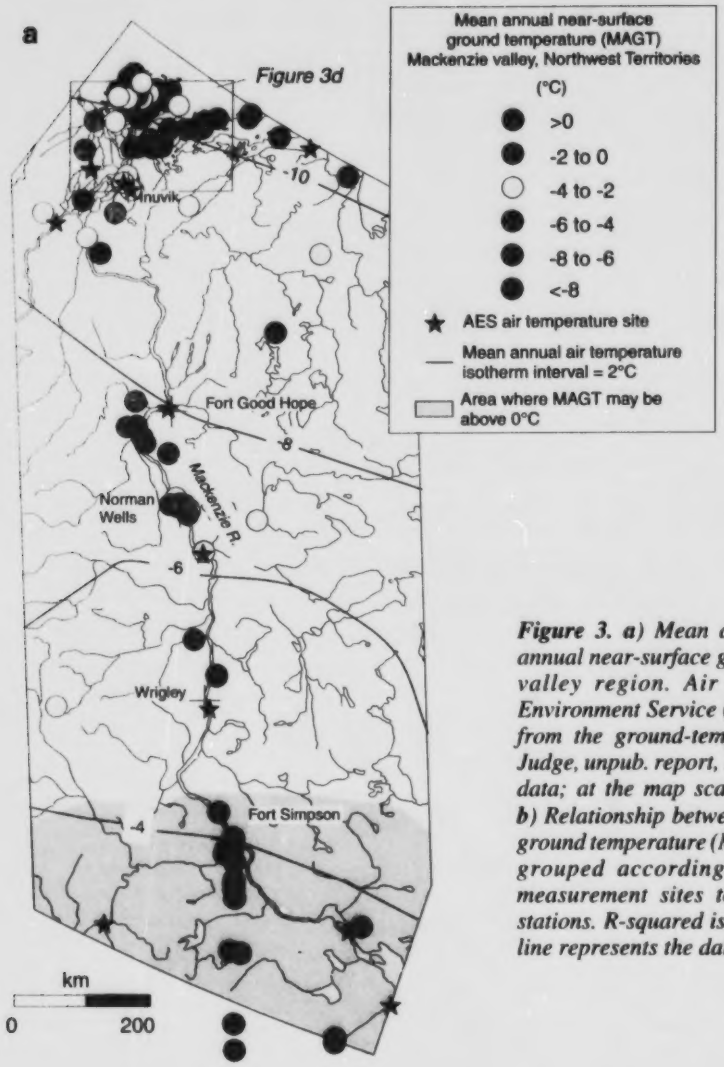
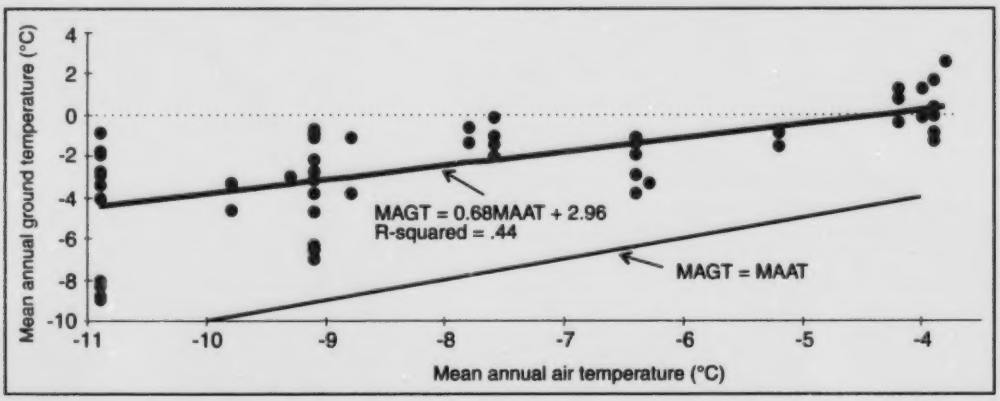


Figure 3. a) *Mean annual air temperature (isotherms) and mean annual near-surface ground temperatures (MAGT) for the Mackenzie valley region. Air temperature data are from Atmospheric Environment Service (1982). Ground-temperature data are extracted from the ground-temperature database (updated from Young and Judge, unpub. report, 1985). Note: there are over 175 sites with MAGT data; at the map scale one dot frequently represents several sites.* ***b)*** *Relationship between mean annual air (MAAT) and mean annual ground temperature (MAGT) for the Mackenzie valley. Data points are grouped according to the proximity of ground-temperature measurement sites to Atmospheric Environment Service climate stations. R-squared is a numerical measure of how well the equation line represents the data. A value of 1 would be a perfect fit.*



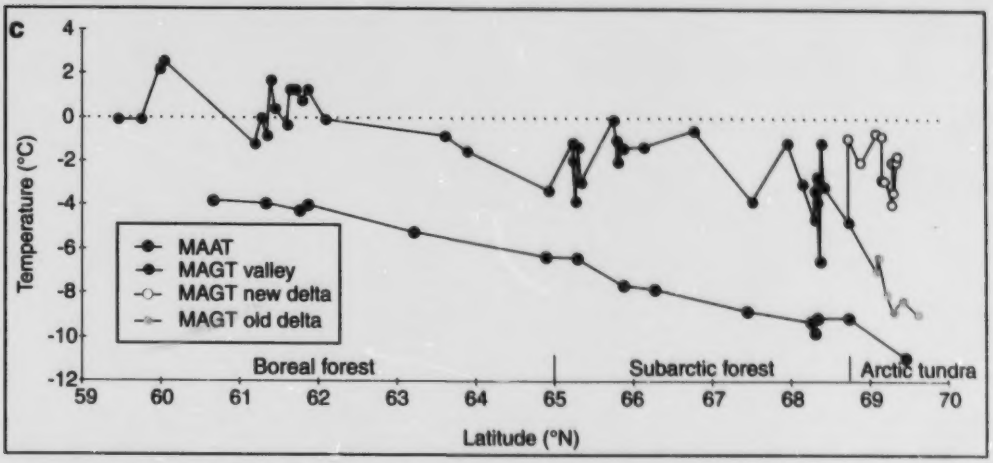


Figure 3c) *The variation in mean annual air (MAAT) and ground temperature (MAGT) along the Mackenzie valley. Ecoclimatic regions (Ecoregions Working Group, 1989) are indicated along the latitude axis. In the Mackenzie Delta area, ground temperatures contrast markedly between the modern (new) delta and the adjacent tundra uplands (old delta). Note: points are joined with lines to highlight differences between each category; the lines are not necessarily representative of interpolated temperatures.*

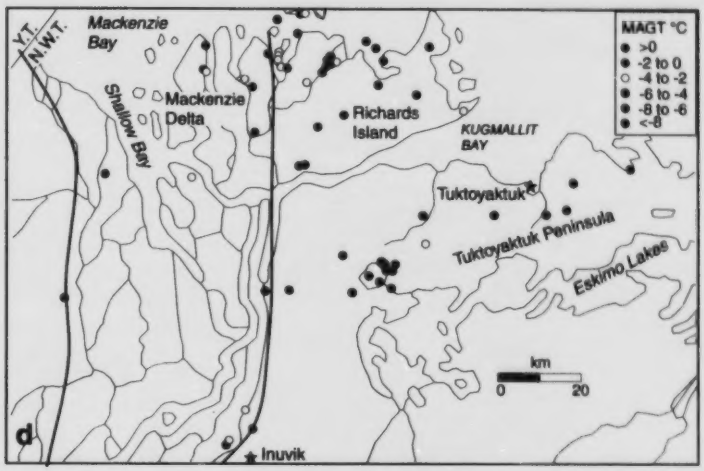


Figure 3d) *Enlargement of part of the Mackenzie Delta area denoted on Figure 3a. Mean annual ground surface temperatures (MAGT) on the Mackenzie Delta are generally warmer than the adjacent tundra uplands of Richards Island and Tuktoyaktuk Peninsula.*

around Fort Simpson, for example, the mean annual air temperature is approximately -4°C, but near-surface ground temperatures are between 2 and -2°C. The influence of local factors can also be seen in the latitudinal profile of air and ground temperature in Figure 3c. There is a decrease in both mean annual air and ground temperature in the Mackenzie valley at higher latitude, but there is greater variation in ground temperature. There is an exception to this general trend in the western Mackenzie Delta region (*see* Fig. 3d). On the modern delta, ground temperatures are considerably warmer than adjacent tundra uplands. In fact, ground temperature here is generally warmer than -4°C, which is similar to that observed in the Mackenzie valley hundreds of kilometres to the south. Ground temperatures are about -8°C at a similar latitude on Richards Island. Vegetation (discussed later with specific examples) which leads to deeper snow cover, and the large amount of water cover maintained by lakes and river channels, are the main factors contributing towards maintaining the warmer ground temperatures in the Mackenzie Delta.

SHALLOW GROUND THERMAL REGIME

Over the past 30 years, many boreholes have been instrumented in the Mackenzie valley for measurements of the shallow ground thermal regime. These boreholes were generally drilled to obtain geotechnical and geothermal data on soils, and to determine permafrost distribution for engineering design or scientific research purposes. In the mid 1970s, analyses of both the shallow and deep thermal regime of the Mackenzie valley were published by Judge (1973, 1975);

these studies included discussion of heat flow, thermal properties, permafrost thickness and temperatures, and annual ground thermal regimes. Since 1980, many new boreholes and studies have contributed to the knowledge and baseline information on the shallow thermal regime in the Mackenzie valley, for example, those related to the Norman Wells pipeline (Pilon et al., 1989; MacInnes et al., 1989, 1990). Figure 4 shows ground-temperature envelopes from a selection of the boreholes throughout the Mackenzie valley and provides subsurface information to complement the near-surface data shown on Figure 3. These plots, which are with one exception from mineral soils, duplicate the general relationship, outlined in the preceding section, between ground temperatures and latitude along the Mackenzie valley.

The temperature profiles at the southern end of the Mackenzie valley (Fort Simpson region and south) show the variability which exists within the discontinuous permafrost zone. In Figure 4, ground-temperature profiles from this area show both an unfrozen mineral site (84-4A) and a frozen mineral site (85-8A), as well as a permafrost site in organic terrain (85-12B). Permafrost occurrence in this region is strongly linked to the presence and nature of peatlands (*see* Aylsworth and Kettles, 2000). Locally, throughout the Mackenzie valley, differences in several factors modify the ground thermal regime. Examples of many of the local influences and the resultant variability in the ground thermal regime are presented below.

Elevation

Increasing elevation can be expected to result in decreasing mean annual air temperatures. Thus, at the southern end of the Mackenzie valley, along the Norman Wells pipeline, permafrost occurrence increases in a southerly direction when terrain rises onto the Alberta Plateau (*see* Burgess and Lawrence, 2000). This relationship with altitude, however, does not always prevail. Figure 5 shows the temperature envelopes for two sites near Norman Wells. The Canyon Creek site is located on the floor of the Mackenzie valley, while the Kee Scarp site is located approximately 250 m higher in elevation and about 20 km to the north. The ground temperatures are warmer at the higher elevation, and permafrost is absent. By contrast, permafrost exists in the valley floor. Such a difference near Norman Wells may be related to the occurrence of pronounced winter air-temperature inversions in the region (*see* Dyke (2000a) for a discussion of inversions). The curvature in the profile at the Kee Scarp site also suggests that there has been recent ground-surface warming, whereas there is no evidence of this at the Canyon Creek site (*see* Dyke, 2000a). These differences emphasize the local variability in climate and ground thermal regime, and also suggest that the local response to any climate warming may be affected by the presence of inversions. There may also be differences in snow cover between the two locations which may contribute to their different response.

Bedrock versus mineral soil

The depth of penetration of the annual temperature wave (i.e. depth of zero annual amplitude) and the damping of the amplitude vary according to the nature of the subsurface materials. In bedrock, with little or no soil cover, there is less rapid attenuation of the annual wave. Thus, at the Kee Scarp site (Fig. 5), where a mere 20 cm of organic material and soil overlie bedrock, the depth of zero annual amplitude is about 20 m. At Canyon Creek, where bedrock is overlain by about 9 m of coarse till which is in turn covered by 40 cm of peat, the depth of zero annual amplitude is only about 10 m. For a given depth, the amplitude of the annual wave is greater in bedrock. For example, temperatures at 2.5 m depth have a range of 5°C at Kee Scarp, compared to 1°C at a similar depth at Canyon Creek. The active layer at the Canyon Creek site is about 1.5 m, whereas the frost penetration layer (the layer that freezes and thaws each year where there is no permafrost) at the bedrock site is around 3.5 m (based on 1987–1988 data).

Drainage

Another borehole at Canyon Creek and a borehole at Norman Wells (but this time in the valley floor) are used to show the influence of drainage on ground temperature and are compared in Figure 6. The soils at the Norman Wells borehole are lacustrine in origin and consist of 6 m of fine-grained silty clay overlying shale; the soils at Canyon Creek are from a coarse-textured ground moraine with 9 m of silty clay overlying shale. Both sites have a surface layer of around 40 cm of peat, and black spruce cover; however, the ground surface is hummocky and poorly drained at Norman Wells, whereas it is flat and well drained at Canyon Creek. Ground-ice contents are very low in the till and higher in the lacustrine clay. Ground temperatures are over 1°C warmer at Canyon Creek, and permafrost thickness there is 25 m. At the Norman Wells site, permafrost thickness is estimated (based on extrapolation of the geothermal gradient) to be about 55 m. The active layer at Norman Wells is about 50 cm, compared to 1.5–2 m at Canyon Creek.

Slope aspect

The orientation of a slope, or its aspect, is one factor determining the amount of incoming solar radiation received at the ground surface. The temperature envelopes plotted from two boreholes on the slopes on either side of Canyon Creek (Fig. 7) reveal that the west-facing slope, which receives greater solar radiation, experiences warmer conditions in the top 15 m than does the east-facing slope. Snow depths also differ on either side of the creek. Snow data from several years in the late 1980s (Burgess, 1993) revealed that snow depths are generally greater at the west-facing site than at the east-facing site. This may be related to differences in wind conditions, as well as to differences in site vegetation. The west-facing slope has a somewhat open cover of black spruce, whereas the east-facing slope has a denser cover of white spruce. More snow would thus be intercepted by the canopy at the latter site, and this could contribute to the colder winter minimum temperatures observed there. Colluvium is

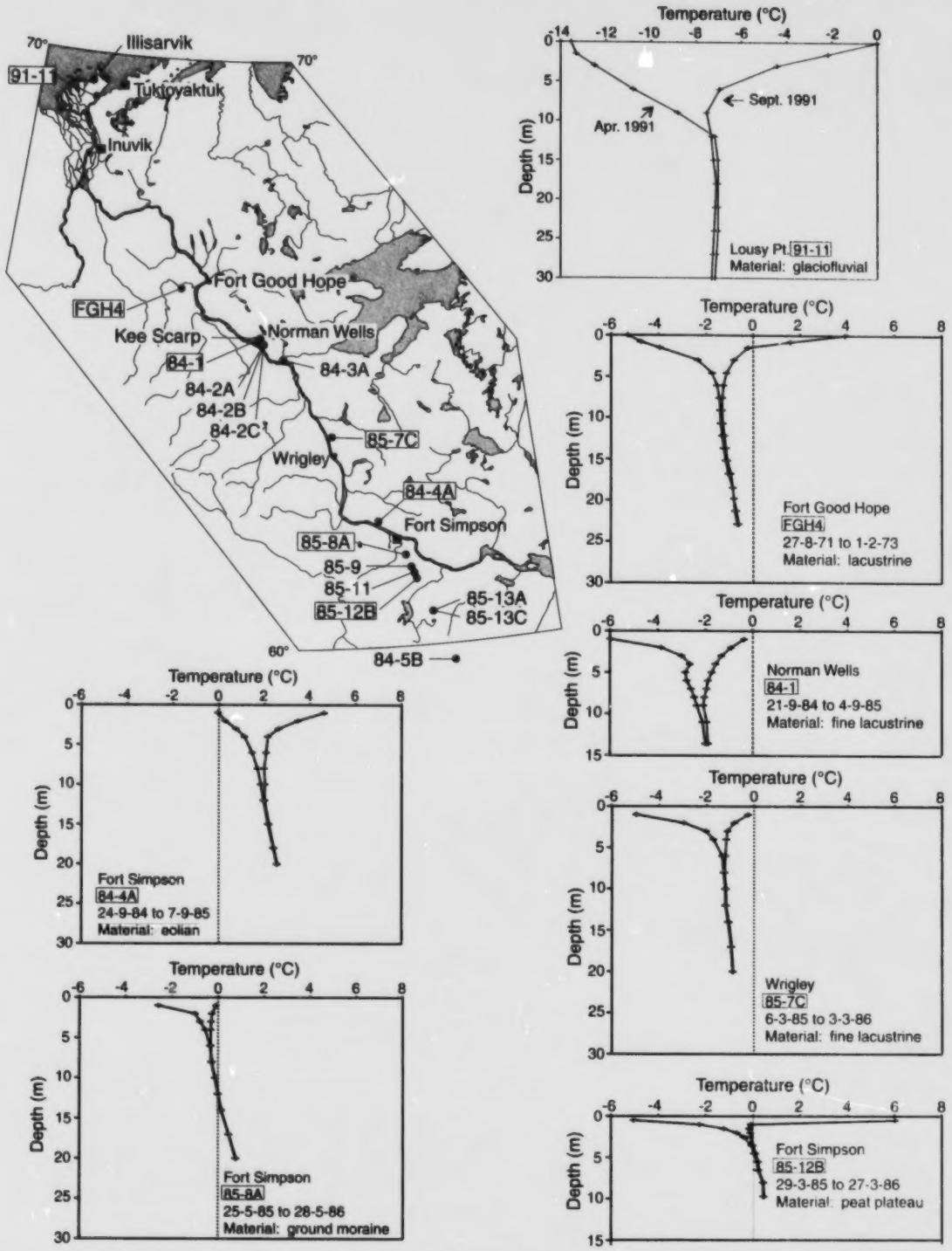


Figure 4. *Ground-temperature envelopes from boreholes in mineral terrain (with exception of site 85-12B which is in organic terrain). Borehole locations are labelled on the map. Other labelled sites are discussed elsewhere in the paper. Material and dates of data collection are indicated on each temperature-envelope plot. Temperature envelopes are based on a series of up to 12 temperature logs recorded over a period of approximately 1 year; with the exception of site 91-11, where the envelope is based on two logs, each of which is individually labelled.*

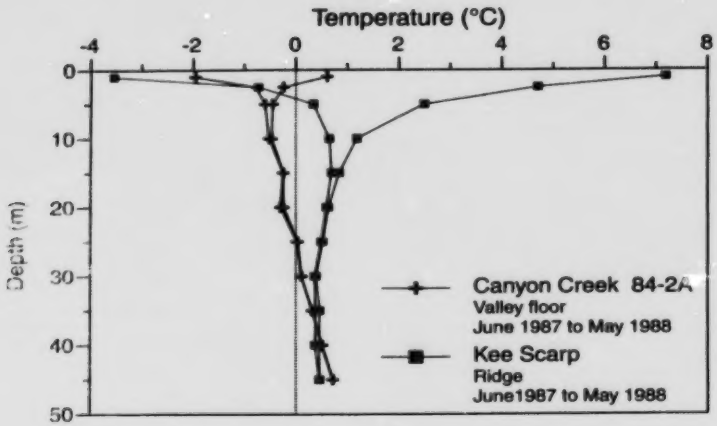


Figure 5.

Maximum and minimum ground temperatures recorded at two sites, located 20 km apart, near Norman Wells over the twelve-month period from June 1987 to May 1988. The Canyon Creek borehole is located on level terrain in the Mackenzie valley floor (site 84-2A), while Kee Scarp is located on a ridge some 250 m higher in elevation.

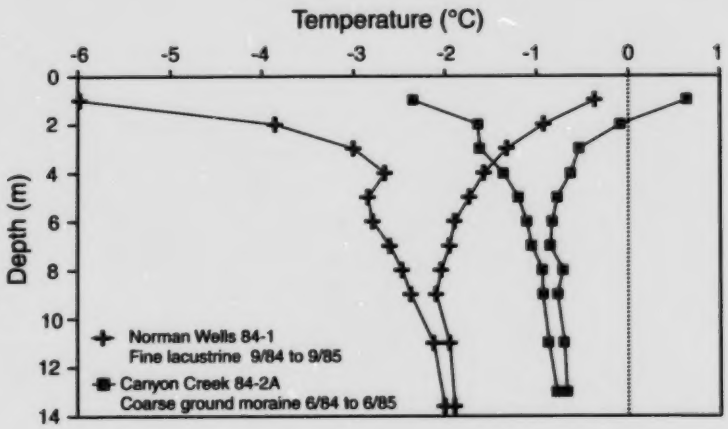


Figure 6.

Ground-temperature envelopes from two mineral soil sites on level terrain in the valley bottom: a fine lacustrine site at Norman Wells (84-1-T4) and a coarse ground moraine 20 km away at Canyon Creek (84-2A-T4). Data for the former site are from the twelve-month period from September 1984 to September 1985, while for the latter site the period is from June 1984 to June 1985.

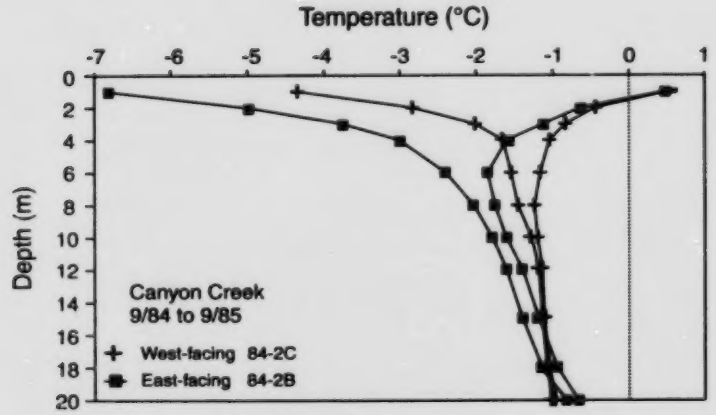


Figure 7.

Ground-temperature envelopes from adjacent opposite-facing slopes at Canyon Creek, for the twelve-month period from September 1984 to September 1985: site 84-2B-T4 is on the east-facing slope and site 84-2C-T4 is on the west-facing slope.

thinner at the east-facing site (with bedrock less than 3 m from the surface), whereas bedrock is at greater than 6 m depth at the west-facing site. These differences in lithology would explain the different geothermal gradients observed at the sites in Figure 7.

Borehole temperatures from the discontinuous permafrost zone

Mineral soil and insulating organic-layer variability

Figure 8 shows the variability in permafrost conditions in mineral soils in the vicinity of Fort Simpson, within the discontinuous permafrost zone. In this area, the presence of a surface organic layer is one of the key factors necessary for the existence of permafrost. Peat changes thermal conductivity during the year, becoming a particularly good insulator (low conductivity) in the summer when it is dry, and a good conductor in the fall and winter when it is wet. This characteristic enables permafrost to be preserved in an otherwise nonpermafrost environment. Site 84-4A is located in sandy soil at the base of a sand dune in eolian terrain east of the Mackenzie River and north of Fort Simpson. Soils at site 85-8A consist of 5 m of sand and silt overlying ice-rich clay and covered by 0.7 m of peat. At site 85-9, coarse granular soils have essentially no insulating surface organic layer, whereas at site 85-11, a 30 cm cover of peat overlies 1–2 m of gravel and sand which, in turn, rest on clay. Sites 84-4A and 85-9 with no surface peat have no permafrost and mean ground temperatures are around 1.5°C. Sites 85-8A and 85-11, both with a layer of peat at the surface, have warm permafrost with temperatures greater than -1°C at the former site, and warmer than -0.5°C at the latter. Permafrost thickness is also greater at 85-8A (11 m) than at 85-11 (2 m).

Peatland types: fen versus peat plateau

Peatlands, which are formed by the accumulation of plant materials, are organic wetlands that have 40 cm or more peat. In the southern parts of the Mackenzie valley, peatlands cover a large percentage of the terrain. The dominant forms of peatland are fens, which are unfrozen, and peat plateau bogs, which are frozen. A discussion of the characteristics and distribution of peatlands in the Mackenzie valley is given in Aylsworth and Kettles (2000). Peat plateaus may also contain internal collapse features which are unfrozen. Figure 9 illustrates the difference in fen and peat-plateau ground thermal regimes recorded at two peatland sites about 75 km north of the Alberta–Northwest Territories border. One site is located on a frozen peat plateau, with 1.6 m of peat and open black spruce forest, while the other is located 400 m to the south in an adjacent unfrozen patterned fen, with 5 m of peat and sedge with some dwarf birch and stunted black spruce. The active layer in the peat plateau bog is less than 1 m and permafrost (at temperatures less than -0.5°C) extends to a depth of about 4 m. Ground temperatures in the fen are about 2 degrees warmer, and seasonal frost penetration is about 1 m.

Unfrozen mineral versus frozen organic soil

The influence of a peat layer on ground temperatures is also shown in Figure 10. It compares frozen organic terrain of a 2 m high peat plateau some 70 km south of Fort Simpson (site 85-12B) to unfrozen mineral terrain some 25 km to the north (site 85-9) with no surface peat layer. Permafrost is only a few metres thick at the peat plateau site and at warm temperatures within 0.5° of 0°C. There is no permafrost, and ground temperatures are about 2°C warmer at the mineral soil site.

Influence of bodies of water

Ground temperatures are also strongly moderated by the presence of bodies of water. Under large and deep water bodies such as the Mackenzie River or large lakes, permafrost may be completely absent. The proximity to water bodies will also frequently produce a warming influence on ground temperatures of the surrounding shores or banks. Smaller bodies of shallow water, such as smaller lakes where bottom waters do not freeze in the winter, often are underlain by an unfrozen zone, or talik, which lies above the permafrost. Figure 11 shows a temperature cross-section beneath Illisarvik Lake on Richards Island in the Mackenzie Delta (*from* Burgess et al., 1982) and highlights the talik which existed to depths of almost 30 m beneath the centre of the lake prior its drainage in 1978. Illisarvik Lake was artificially drained as part an experiment to study permafrost aggradation in drained-lake

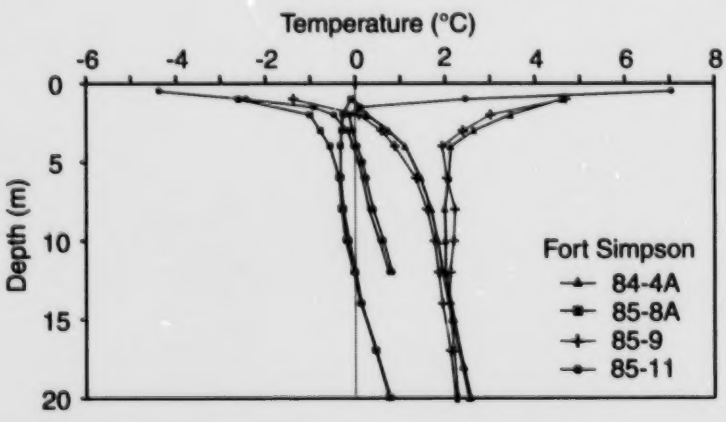


Figure 8.

The 1986 ground-temperature envelopes from several mineral soil sites in the Fort Simpson region. Soils at sites 85-9 and 85-11 are coarse granular gravels or sands, whereas at sites 84-4A and 85-8A soils are finer textured sands and silts. Permafrost exists at 85-8A and 85-11 due to the presence of a surface organic layer 30 cm or more in thickness. At sites 84-4A and 85-9 there is essentially no insulating surface organic layer.

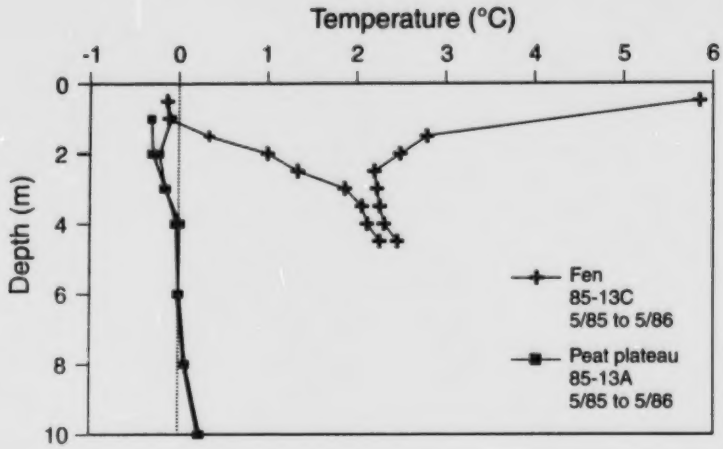


Figure 9.

Temperature envelopes from a peat plateau (site 85-13A) and a nearby fen (site 85-13C) along the Norman Wells pipeline; data from twelve-month period from May 1985 to May 1986.

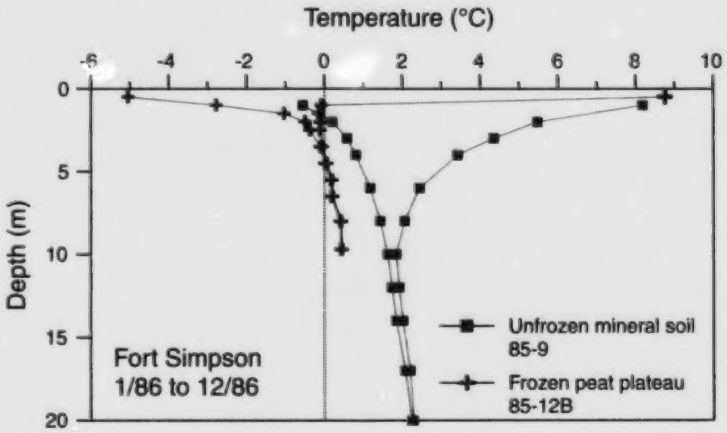


Figure 10.

Comparison of two sites south of Fort Simpson: a peat plateau with permafrost (site 85-12B) and unfrozen mineral soil (site 85-9): 1986 ground-temperature envelopes.

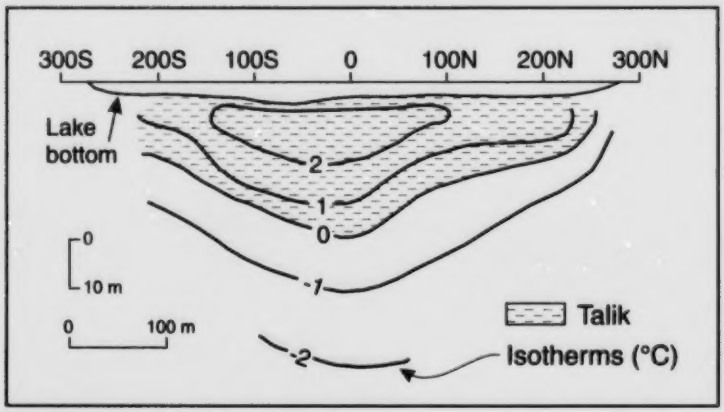


Figure 11.

Temperature cross-section beneath Illisarvik Lake, Richards Island, Mackenzie Delta, prior to its drainage in 1978.

bottoms (Mackay, 1981). Along the actively eroding coasts and shorelines of the Mackenzie valley, catastrophic lake drainage is a common geomorphic process, and has often resulted in the formation of pingos, as permafrost aggrades into the exposed unfrozen lake bottom sediments.

In addition to lake drainage, parts of the Mackenzie River channel, Mackenzie Delta front and offshore experience continual changes as river channels migrate and shorelines and bars shift. These dynamic processes lead to corresponding dynamic and diverse vegetation, snow cover, and ground thermal regimes, as discussed in Dyke (2000b; *see also* Smith and Hwang, 1975).

Snow cover

Snow acts as an insulator, and the difference between air and ground surface temperature tends to be greater in areas covered by a thick snow pack (Goodrich, 1982). Data from Canyon Creek, for example (Stuart et al., 1991), shows that as snow depth increases, changes in air temperature are accompanied by smaller changes in soil temperature. Once the snow cover is greater than 50 cm thick, soil temperature is only affected by large sudden changes in air temperature.

The dramatic influence of snow depth on shallow ground temperature is clearly illustrated in Figure 12 with four GSC sites on the modern Mackenzie Delta. Mean annual ground temperatures range from -9°C at a site with almost no snow cover to about -2.5°C at a site with a significant snow cover. Variations in snow cover are related to the different vegetation which exists at the sites. Vegetation is non-existent at the sand flat and snow is usually blown off the site resulting in little snow cover in winter. The shallow ground temperature on the sand bar is therefore lower than at the other sites which each have a vegetation cover and thicker snow packs. At the dwarf willow (tundra) site, the low vegetation cover accumulates about 10 cm of snow. Ground temperatures at this site are therefore lower than at the other vegetated sites. Snow depths at the sedge site increase to about 20 cm. The taller willows trap snow to a depth in excess of 60 cm. Where taller willows occur, snow depths can reach 1 m. Shallow ground temperatures are warmest at the willow-covered site.

RESPONSE TO SURFACE DISTURBANCE

Climate variability or climate change due either to natural or anthropogenic causes may lead to ground-surface temperature changes. In addition, other activities which directly disturb the ground surface, such as clearing of vegetation for roads, pipelines or other rights-of-way, or destruction of the vegetation and surface organic material due to forest fires also affect the ground-surface temperature. This section illustrates changes in the ground thermal regime that have occurred over a ten-year interval in response to the construction of the Norman Wells pipeline.

The Norman Wells pipeline, in operation in the Mackenzie valley since 1985, is 328 mm in diameter and runs from Norman Wells to Zama, Alberta. The pipeline is buried and traverses the discontinuous permafrost zone more or less parallel to the Mackenzie River. Clearing of trees on the right-of-way and construction took place from 1983 to 1985. A series of monitoring sites were established by the government along the route to study the impact of construction and operation on permafrost terrain (Pilon et al., 1989; MacInnes et al., 1989, 1990). The effect of suddenly removing the trees is akin to a sudden step increase in surface temperature; during construction, the surface organic layer was also frequently scraped to obtain a stable working platform. The rate and magnitude of ground-temperature changes have varied according to soil type, ice content, and initial ground thermal condition (Burgess and Riseborough, 1990; Burgess, 1992).

Examples of the variability in the terrain response are illustrated in the cross-sections of Figure 13. These show the depth of thaw (active layer) at the time of site establishment in the 1980s and ten years later in the mid-1990s. Note that thaw depths are greater in the vicinity of the buried pipe due to the warm pipe-operating temperatures (generally >0°C), the increased level of terrain disturbance in the pipeline trench, and the fact that the trench backfill is frequently subsided and wetter than the adjacent right-of-way (ROW). In general, away from the pipe, thaw depths on-right-of-way were initially around 1 m. Increase in thaw depth on the right-of-way due to clearing and construction at these particular sites has varied from 1–2 m at site 84-3A, to 4 m at site 85-7C

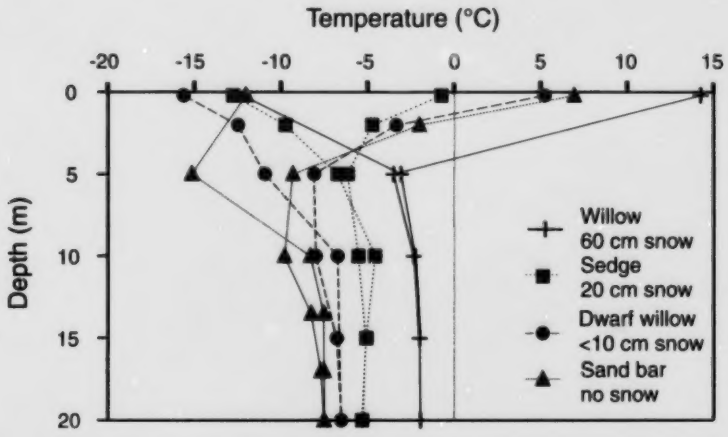


Figure 12.

Ground-temperature profiles at four sites with different vegetation on the outer Mackenzie Delta. The profiles show the effect of different snow depths caused by the height of the respective vegetation. They vary from willows, trapping 60 cm of snow, to an unvegetated sand flat kept clear of snow by the wind. The insulating effect of the snow also reduces the maximum depth of annual temperature variation. Dates of the temperature profiles at each site are March and July 1992.

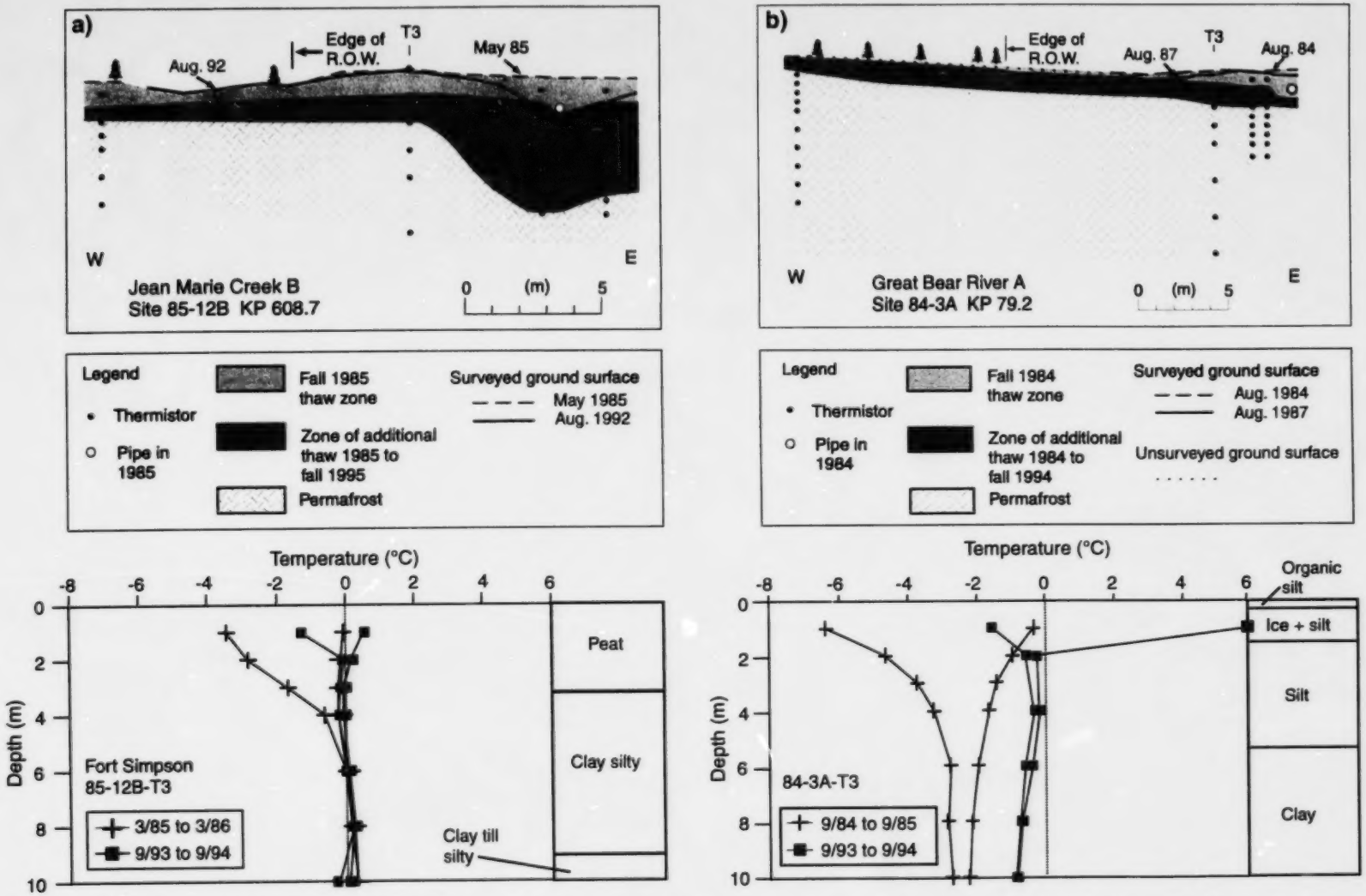


Figure 13. *Cross-sections showing change in thaw depth, permafrost thickness, and surface settlement related to the Norman Wells pipeline. Ground-temperature envelopes show the change in thermal regime on the pipeline right-of-way. Data from several sites illustrate the diversity in the response of different terrain units to surface disturbance.* ***a)*** *Site 85-12B is an ice-rich peat plateau south of Fort Simpson.* ***b)*** *Site 84-3A is on an ice-rich fine-grained alluvial terrace near Tulita (formerly Fort Norman).* ***c)*** *Site 84-5B is also an ice-rich peat plateau on the Alberta Plateau (thicker peat and thicker permafrost than 85-12B).* ***d)*** *Site 85-7C is in fine-grained lacustrine soils north of Wrigley. KP = kilometre post along Norman Wells pipeline.*

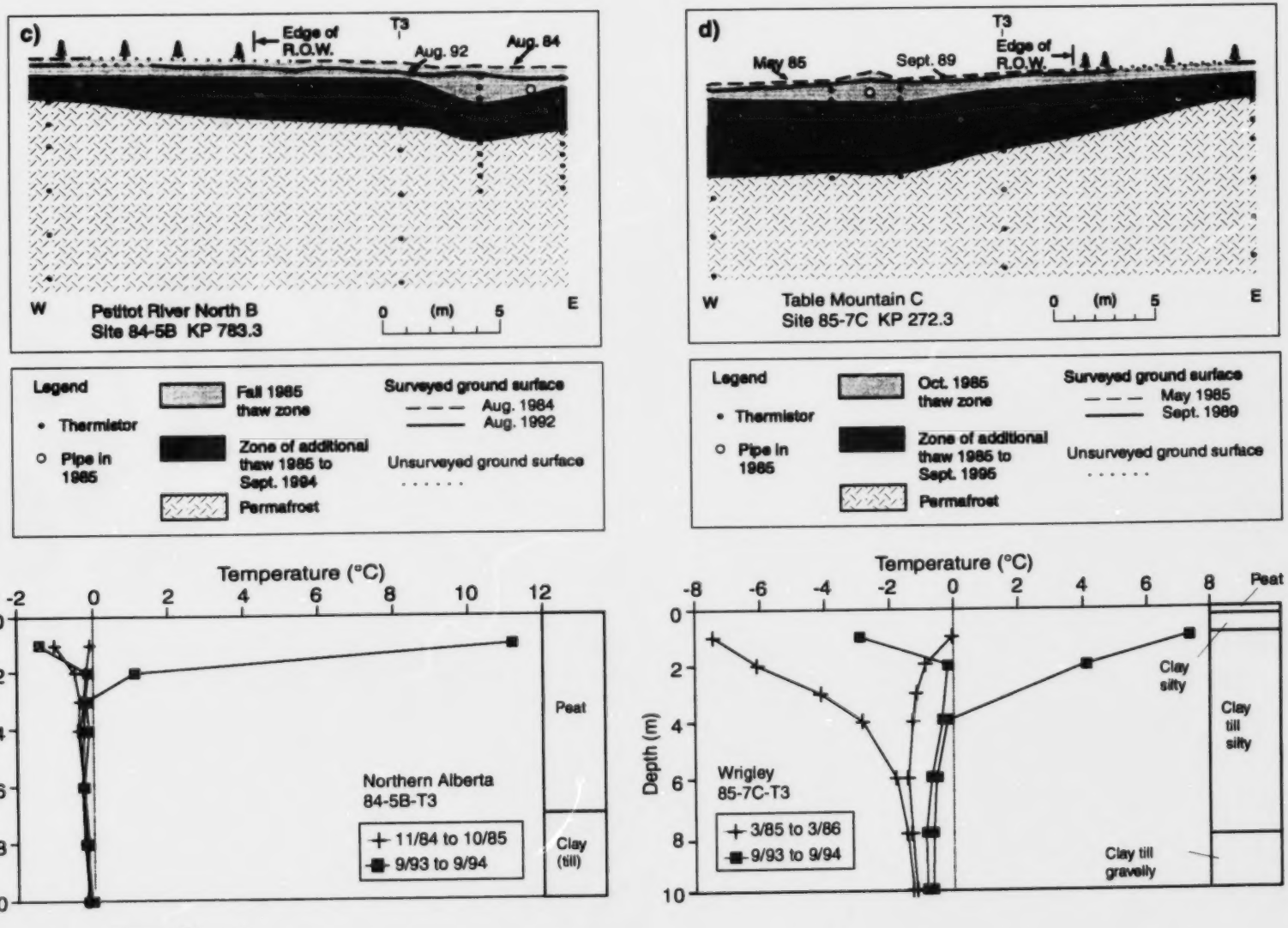


Figure 13 (cont.)

(fine-grained ice-rich lacustrine soils near Wrigley). The smaller increase in thaw depth at site 84-3A is due to a 1 m layer of massive ice near the surface.

The cross-sections are accompanied by annual temperature envelopes on the right-of-way for the first year after site establishment and about 10 years after. Data from the right-of-way cable furthest from the pipe have been selected so as to minimize the thermal effects of the pipe. The two envelopes illustrate the change in the ground thermal regime that has occurred over this time interval. The cross-sections also show the change in surface elevation, i.e. thaw settlement, that has occurred subsequent to site establishment (survey dates vary from site to site). Further discussion of thaw settlement potential for various terrain types can be found in Aylsworth et al. (2000a) and Burgess et al. (2000). Small changes in thaw depth are also observed at the off-right-of-way cable, and these are largely related to the surface disturbance caused by drill-rig movement to the off-right-of-way sites.

In the mid 1990s there have been several large and extensive forest fires in the Mackenzie valley, and these have traversed large sections of the pipeline right-of-way. The response of terrain to forest fires is also site specific and dependant on several factors. Mackay (1995) presented data from a burn area near Inuvik that showed the variability in changes in the active layer depth over time for sites with different initial permafrost and ground-ice conditions, vegetation changes, relief, and snow cover.

The off-right-of-way at pipeline site 84-3A was burnt in the spring of 1995. The existence of the monitoring site will provide both the base-line data and ongoing observations to allow an examination of the impact on the shallow thermal regime. New instrumentation has also been installed on a slope that burnt in the summer of 1994, to examine the thermal impact and recovery of terrain at several elevations along the slope, and to compare these to the conditions at an unburnt location on the slope. The creek valley at this location experienced numerous active-layer detachment slides within weeks of the fire. Slope stability under a warming climate, with or without increased forest fire activity, is a concern in the Mackenzie valley, and is further discussed in Aylsworth et al. (2000b) and Dyke (2000c).

SUMMARY

This paper has presented data that show the broad general relationship, and great local variability, between climate and near-surface ground temperature in the Mackenzie valley. Local variability arises due to differences in environmental and site conditions, such as vegetation, snow cover, elevation, surface organic layer, soil type, moisture content, and thermal properties. Prediction of the response of the shallow ground thermal regime to climate change must thus take into account this variability and the complexity of the interactions between climate above the ground, climate at the ground surface, and climate in the ground.

Geothermal models are frequently employed to assess the ground thermal response to climate change. Many existing models simulate the complete surface energy-balance equation, but frequently there are insufficient data available to characterize all the necessary input parameters. Recently, the use of n-factors (a transfer function between air temperature and temperature at the ground surface which takes into account the influence of vegetation and snow cover) has become more prevalent in geothermal simulations. The characterization and measurement of n-factors for natural surfaces is therefore becoming increasingly important (*see* Taylor, 2000). Smith and Riseborough (1996) presented a new formulation of the permafrost-climate system which links ground temperature to the annual cycle of atmospheric temperature, with the modulating effects of local surface and lithological conditions (n-factors also appear in their formulation).

Assessing the impact of climate warming on the ground thermal regime requires not only a knowledge of the current ground thermal regime, but also a prediction of how air temperatures will change and the availability of a geothermal simulation model which can transfer atmospheric changes in climate to the ground surface. Determination of changes in precipitation and vegetation are also critical. For example, if an air temperature increase is accompanied by a decrease in snow cover, ground temperatures may not increase at all since the latter change can offset the former. Burgess et al. (2000) present one modelling approach that was used to examine various warming scenarios and the possible associated transient changes in permafrost and the shallow ground thermal regime at three locations in the Mackenzie valley.

ACKNOWLEDGMENTS

We would like to acknowledge our colleagues in the Terrain Sciences Division of the GSC for their contribution of data on the ground thermal regime. The collection of the data presented here has in many cases also been the result of collaborative field efforts or funding with other agencies and individuals, notably the Department of Indian and Northern Affairs (Northern Affairs Program) for study sites along the Norman Wells pipeline and in the Mackenzie Delta.

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Deep ground temperatures

A.E. Taylor[1], M.M. Burgess[2], A.S. Judge[3], and V.S. Allen[4]

Taylor, A.E., Burgess, M.M., Judge, A.S., and Allen, V.S., 2000: Deep ground temperatures; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 105–109.*

Abstract: Some 60 holes drilled for hydrocarbon exploration in the Mackenzie valley afford the opportunity to obtain precise measurements of deep permafrost thicknesses (depth to 0°C) and ground temperatures, and are presented here on maps and graphs. Permafrost thickness is less than 100 m in the Mackenzie Delta, and up to 663 m in the adjoining Tuktoyaktuk coastlands. Permafrost varies from 35 m to 143 m in the Norman Wells area, but is thin or absent in the Fort Simpson and Yellowknife areas. Variations in ground temperatures with depth reflect differing thermal properties of soil and rock, proximity to lakes or rivers, and residual thermal disturbance due to drilling. In some cases, subsurface temperatures show features that are relict from several thousand years ago when the local site environment was different from today, due for example, to more widespread lakes, changing shorelines, or river channels.

Résumé : Grâce aux quelque 60 trous forés à des fins d'exploration des hydrocarbures dans la vallée du Mackenzie, on a pu mesurer avec précision l'épaisseur du pergélisol en profondeur (profondeur où est atteinte la température de 0 °C) et les températures du sol, et on a reporté les données sur des cartes et des graphiques. L'épaisseur du pergélisol n'atteint pas 100 m dans le delta du Mackenzie mais elle peut atteindre 663 m dans la région littorale de la péninsule de Tuktoyaktuk. Le pergélisol varie en épaisseur de 35 à 143 m dans la région de Norman Wells, mais il est mince ou absent dans les régions de Fort Simpson et de Yellowknife. Les variations de la température du sol en fonction de la profondeur peuvent être attribuables aux propriétés thermiques des sols et des roches, à la proximité de lacs ou de cours d'eau et à la perturbation thermique résiduelle causée par les forages. Dans certains cas, les températures souterraines indiquent un pergélisol relique datant de plusieurs milliers d'années, formé alors que les conditions environnementales locales étaient différentes de celles d'aujourd'hui à cause. Ces changements peuvent être liés, entre autres, à un plus grand nombre de lacs plus vastes, à l'évolution des berges ou à la migration des chenaux fluviaux.

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Pioneering measurements of deep ground temperatures and permafrost conditions were made in the Norman Wells area by R.A. Hemstock (unpub. company report, Imperial Oil Limited, Calgary, Alberta, 1949), and Garland and Lennox (1962), and in the Mackenzie valley by Mackay (1967). Over the past thirty years, subsurface temperature measurements to depths greater than 125 m have been obtained by the Geological Survey of Canada from numerous, specially adapted resource-exploration holes: some 60 such wells are located in the Mackenzie valley, 40 of which are in the Mackenzie Delta–Tuktoyaktuk coastlands area (Judge, 1973a, b; Taylor et al., 1982). This unique data set provided industry and government with the first precise measurements of permafrost thickness (depth to 0°C) and temperature of permafrost. More recently, analysis of geophysical well logs has provided, at less precision, estimates of the depth to the base of permafrost ice-bonding at many more wells (Judge et al., 1987).

Figure 1 shows typical temperature data taken periodically over eight years at a Mackenzie Delta well. The thermal disturbance due to drilling through porous, frozen rock is apparent, and tens of years may be required to refreeze and cool to undisturbed conditions (Lachenbruch and Brewer, 1959); however, such a series of measurements is used to mathematically extrapolate the estimated equilibrium (or undisturbed) temperatures T_{EQ} that are shown in this paper.

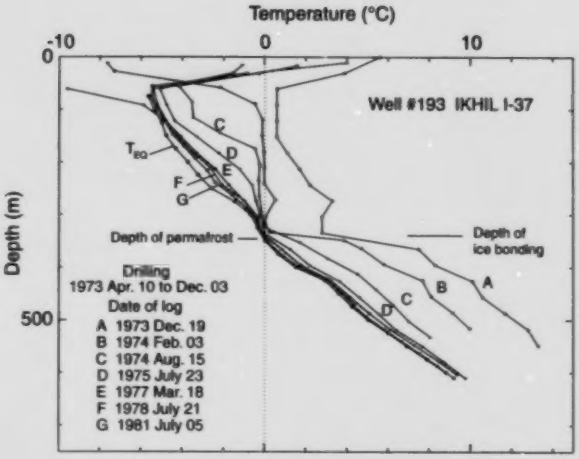

Figure 1. *Precision temperatures measured with a lightweight logging system over several years following drilling of a typical well. As the well cools from the thermal effects of drilling, temperatures 'hang' at the formation freezing point while refreezing takes place, before cooling further towards estimated undisturbed values, T_{EQ}. The depth of the intersection of the equilibrium profile T_{EQ} with 0°C is the depth of permafrost (347 m); the depth of the jump in temperatures in log A taken shortly after drilling is interpreted as the depth of ice bonding (341 ± 8 m). For location, see well #193, Figure 2b.*

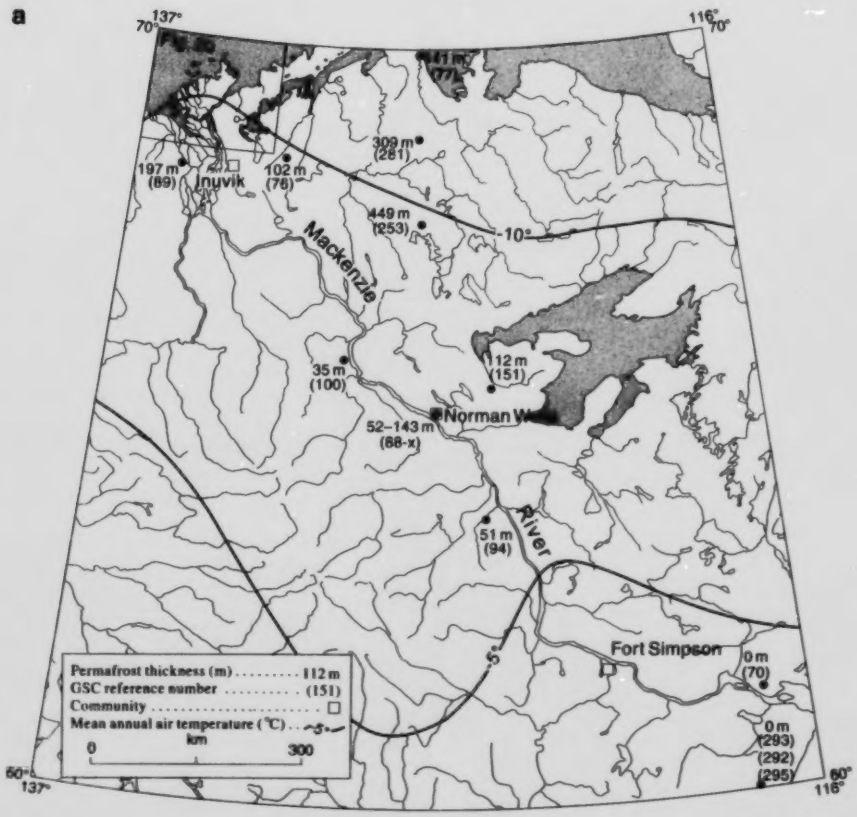

Figure 2.

*Locations of wells where deep ground temperatures have been measured and permafrost thicknesses (m) have been determined. The numerals in brackets are reference numbers for identifying profiles in Figure 3. **a)** Mackenzie valley, **b)** Mackenzie Delta and Tuktoyaktuk coastlands.*

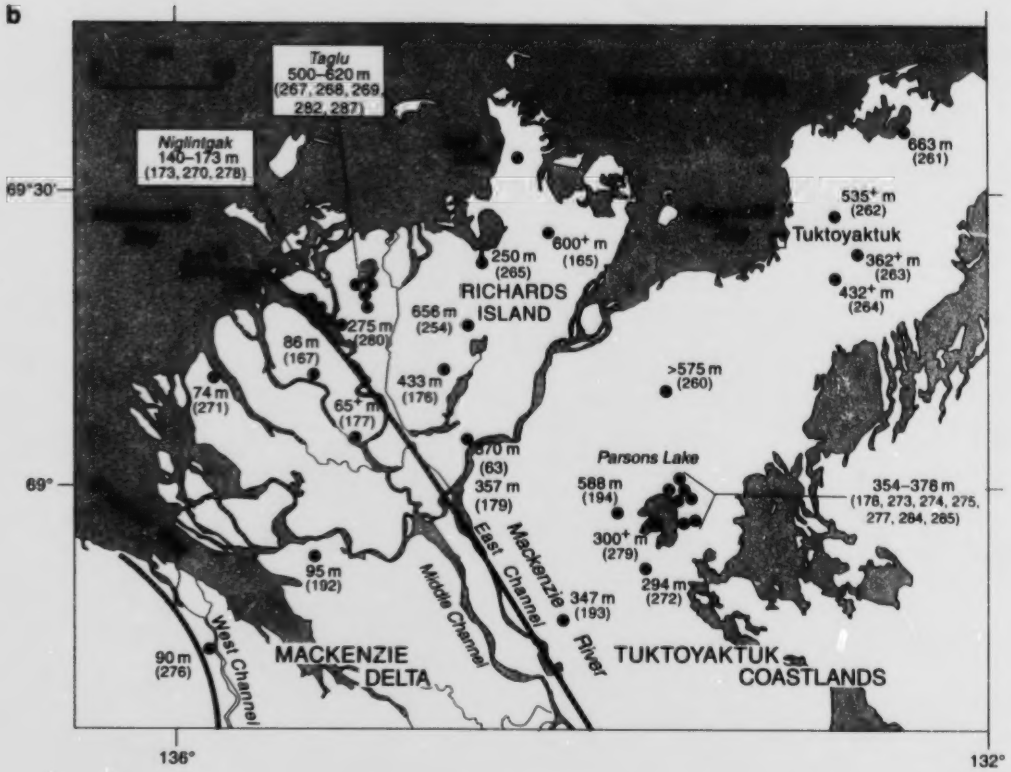

Figure 2. *(cont.)*

Figure 2a is a map of the depth to the base of permafrost determined at these wells. Permafrost is absent or thin in the extreme upper valley near 60°N, varies from 35 m to 143 m in the mid-valley around Norman Wells, and from 65 m to 663 m in the lower valley and Mackenzie Delta.

Permafrost thickness is a function of surface (air) temperatures for the past hundreds to tens of thousands of years, and of ground thermal properties. Permafrost is less than 100 m thick in the modern Mackenzie Delta (Fig. 2b), which was a deep marine trough prior to the formation of the delta in the past several thousand years; modelling suggests that about 3000 years of exposure to arctic air temperatures were required to grow the 86 m of permafrost observed at well #167 (Taylor et al., 1996). In contrast, immediately to the east, permafrost is up to 663 m thick in the Tuktoyaktuk coastlands, which was land exposed to severe temperatures for most of the past 100 000 years. In a small region, permafrost may be twice as thick in sands or sandstone as in clays or shale, because the thermal conductivity of sands and sandstone are typically twice that of clays or shale: heat loss under severe arctic conditions is greater in the higher conductivity environment.

A set of graphs, illustrates the variety of permafrost temperatures observed in the Mackenzie valley (Fig. 3). Thermal conductivity accounts for most of the variation in temperature with depth, as discussed above. The increase of temperature with depth is greater in the shale and limestone of the Mackenzie fold belt (Fig. 3b, c) than in deltaic sands and gravels of the Tuktoyaktuk coastlands (Fig. 3e). Other variations may be attributed to proximity to lakes or rivers, and to past changes in surface temperature. For instance, in the Taglu area, widespread lake or sea encroachment for a period of a few thousand years during the Holocene occurred at sites that are now on land, and temperatures remain close to the freezing point over a large depth interval because of this (Fig. 3h; Taylor et al., 1996).

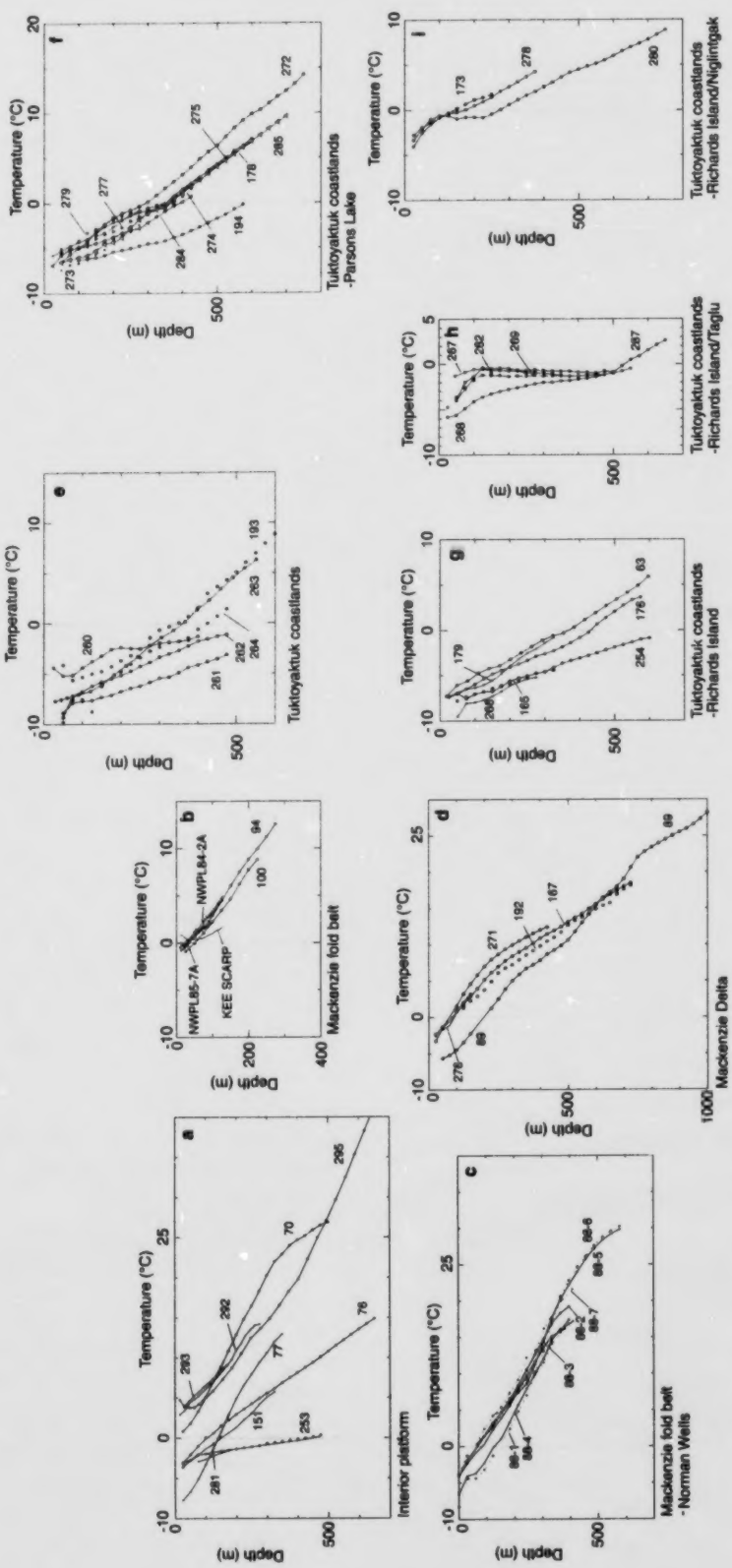

Figure 3. *Profiles of calculated equilibrium temperatures versus depth, grouped by convenient physiographic area; Niglingak and Taglu are hydrocarbon exploration fields on western Richards Island (Fig. 2b). Numbers beside each profile provide cross-reference to Figure 2. The original data and further details are available in Taylor et al. (1982).*

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Indian and Northern Affairs Canada, the Canada Oil and Gas Lands Administration, the National Energy Board and various petroleum exploration companies co-operated in the preservation of these wells to enable extended temperature observations. Logistic support was provided by the Polar Continental Shelf Project and by Energy, Mines and Resources Canada. T. Lewis is thanked for critically reviewing the manuscript.

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Relationship of ground temperatures to air temperatures in forests

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Taylor, A.E., 2000: Relationship of ground temperatures to air temperatures in forests; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 111–117.*

Abstract: Ground temperatures differ from air temperatures because of effects of vegetation and humus, other site characteristics, and snow cover. For engineering work, ground surface temperatures may be estimated from the more readily available air temperatures through the use of a single parameter, the n-factor, to represent these effects. In the Mackenzie valley, n-factors have been calculated from measurements of air and ground temperatures at 27 diverse, natural sites. Generally, the n-factor for the thawing (spring-summer) season is higher in more open areas than in shady forests with thick moss. At a particular site, the n-factor for the freezing (fall-winter) season is usually lower than the corresponding n-factor for the thaw season, because of the reduced impact of air temperatures and solar radiation on the ground due to snow cover. One application of n-factors is the calculation of the depth of soil that thaws each year, called the active layer.

Résumé : Les températures du sol diffèrent des températures de l'air en raison de différents éléments, que ce soit la présence de végétation et d'humus, l'existence d'un manteau nival, etc. Dans les travaux d'ingénierie, on peut estimer les températures de la surface du sol à partir des températures de l'air disponibles en utilisant un seul paramètre, le facteur n, pour représenter l'effet de ces éléments. Dans la vallée du Mackenzie, on a calculé les facteurs n à partir des températures de l'air et du sol enregistrées à 27 sites naturels. En général, le facteur n pendant la saison de dégel (printemps et été) est plus élevé dans les zones dégagées que dans les forêts ombragées où le sol est recouvert d'une mousse épaisse. En un endroit donné, le facteur n est habituellement plus bas pendant la saison de gel (automne et hiver) que pendant la saison de dégel du fait que le manteau nival réduit les effets de la température de l'air et du rayonnement solaire sur le sol. Le facteur n sert également à calculer la profondeur du mollisol, le sol de surface qui dégèle à chaque année.

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INTRODUCTION

In permafrost areas, the fundamental condition that controls permafrost is the surface temperature. To predict the distribution of permafrost over an area (*see* Wright et al., 2000), it is necessary to know how much effect the snow, vegetation, humus, and standing water have in reducing the impact of air temperatures and radiation on ground-surface temperatures (Fig. 1). For engineering calculations, the effect of these parameters may be considered through a single parameter, the n-factor, that may be readily obtained from air and corresponding ground temperatures in type areas (e.g. Lunardini, 1981). This simplifying concept offers a practical way of using the more readily available air temperatures to predict ground temperatures where the latter are unknown.

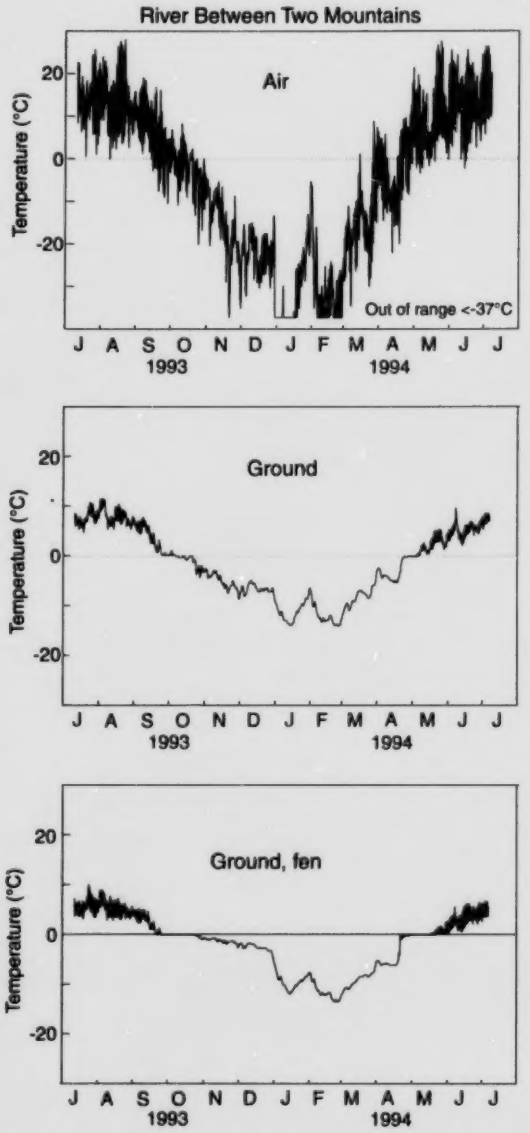

Figure 1. *Typical record of air temperatures (upper graph) and ground temperatures (depth 3–7 cm, middle graph) at an undisturbed forest site (site 'A', Fig. 6). The bottom graph shows ground temperatures in a nearby fen (site 'R', Fig. 6). Delays in freezing in the fall, and in thawing in the spring (zero-curtain effect) observed in the ground temperature arise from the dissipation of latent heat of phase change.*

The thawing season n-factor, n_{th}, is defined as the ratio of the ground surface index of seasonal thawing ($I_{s,th}$) to the air index of seasonal thawing ($I_{a,th}$):

$$n_{th} = \frac{I_{s,th}}{I_{a,th}}$$

and similarly the freezing season n-factor is the ratio of the ground surface index of seasonal freezing ($I_{s,\,fr}$) to the air index of seasonal freezing ($I_{a,\,fr}$):

$$n_{fr} = \frac{I_{s,fr}}{I_{a,fr}}$$

These indices are expressed in degree-days relative to 0°C. Figure 2 shows schematically how n-factors at a site may be calculated from a time-series of air and ground temperatures. An n-factor equal to one results if there is no attenuation (and hence ground-surface temperatures are equal to air temperatures), and the n-factor is less than one for attenuation due to vegetation and snow cover. Note that n_{th} may be greater than one due to direct solar radiation on surfaces devoid of vegetation.

The Geological Survey of Canada has established a latitudinal transect along the Mackenzie valley (Fig. 3) using inexpensive miniature data loggers to record air and ground temperatures that allow the calculation of n-factors (Nixon and Taylor, 1994; Taylor, 1995). Air temperatures are measured every 5 hours in a radiation shield and ground temperatures are measured within 3–7 cm of the surface, generally within a dense peat (e.g. Fig. 1). Figures 4–8 show photographs of a selection of sites chosen to demonstrate the variation in n-factor values that can be readily attributed to site geology and vegetation. Ground-surface temperatures of sites designated "A" are measured in the immediate vicinity of air temperature measurements in radiation shields. Ground-surface temperatures at sites designated "R" are measured at remote sites which are usually within 100 m of the "A" site. It must be emphasized that these n-factor values are for a particular year of data; values will vary somewhat from year to year, largely due to annual variation in snow cover.

The n-factors for the thaw season appear to be highest in more open areas having good sun exposure, and lower in shady forests with thick moss. For example, Figure 4, Mackenzie Highway kilometre post (kp) 504.7: n_{th} = 0.30 in a shady upland hardwood and spruce forest with 10 cm or more moss (site 'A'), whereas some 30 m away, n_{th} = 0.79 in mineral soil on an overgrown, but more exposed cutline through similar forest (site 'R'). Figure 5, Mackenzie Highway kp 507.6: n_{th} = 0.66 in a small, exposed bog between a fen and upland mixed forest; the bog was wet and the data logger was submerged at time of

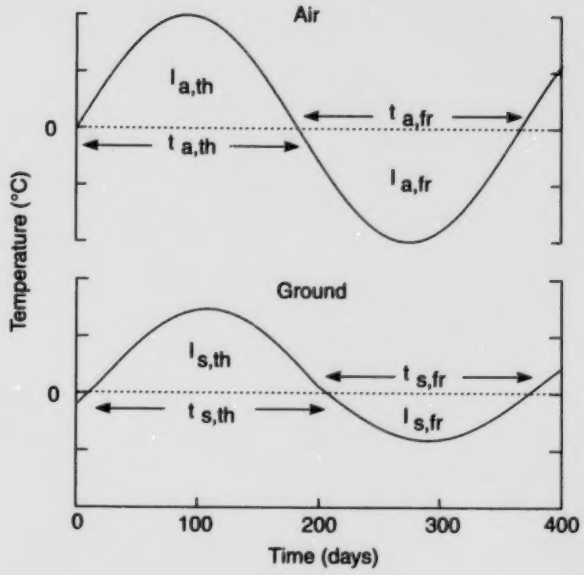

Figure 2. *Schematic of air and ground temperatures for just over one year, showing the seasonal thawing and freezing indices needed to calculate n-factors (see text for definitions). For instance, $I_{a,th}$ is the thawing index (air) and is the area (in degree-days) between the air temperature curve above freezing and the freezing point. The terms $t_{a,th}$ and $t_{a,fr}$ are the length of the thawing season and freezing season, repectively, in air. The terms $t_{s,\ th}$ and $t_{s,\ fr}$ are the length of the thawing season and freezing season, respectively, at the ground surface. Note that the ground-temperature variation lags the air temperatures due to the time required for the signal to penetrate the vegetation and snow.*

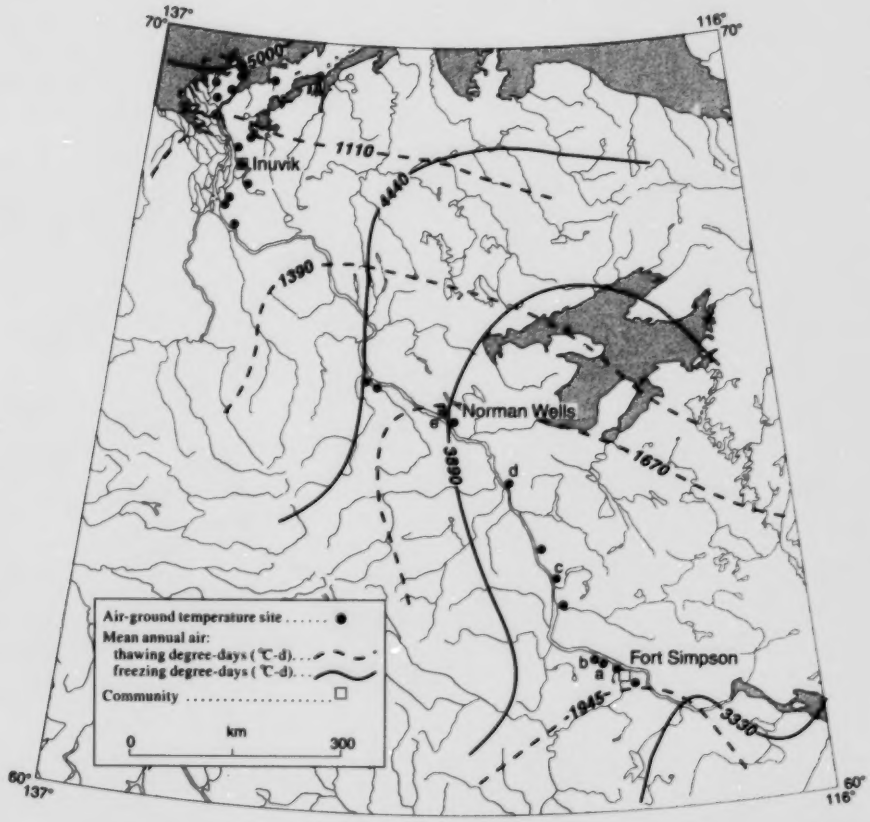

Figure 3. *Location of sites where air and ground temperatures are being recorded for the calculation of n-factors. Labels a to e reference sites shown in Figures 4–8, respectively. Contours are shown for mean annual thawing ($I_{a,th}$) and freezing indices ($I_{a,fr}$) in air (degree Celsius-days above and below 0°C, respectively).*

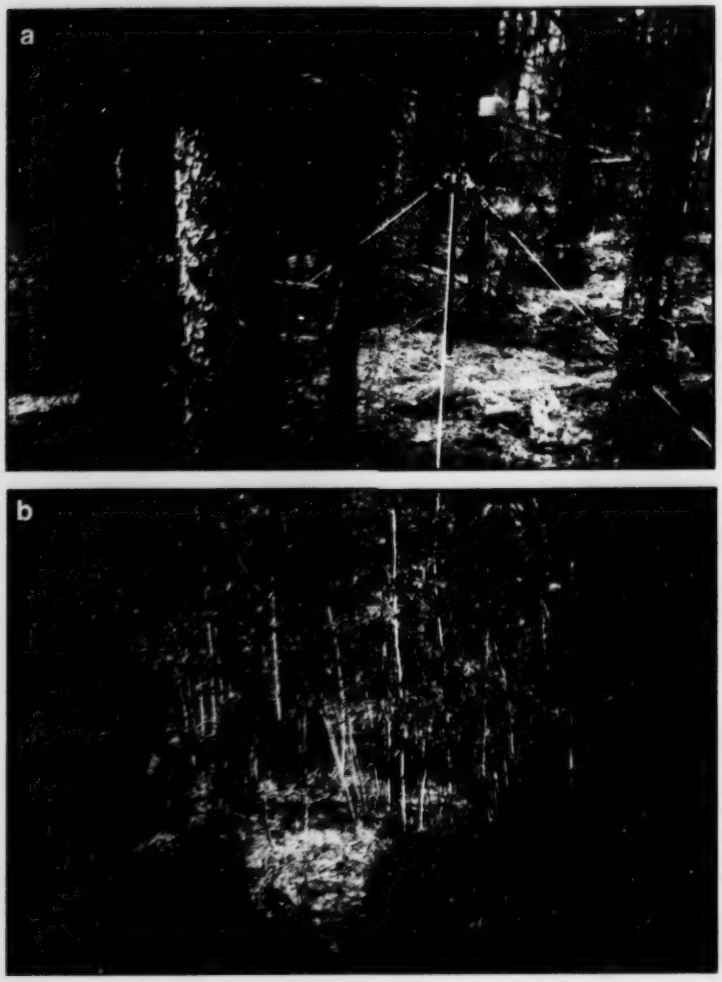

Figure 4.

Mackenzie Hwy. kp 504.7
Site A: n_{th} = 0.30 n_{fr} = 0.20
Site R: n_{th} = 0.79 n_{fr} = 0.17
***Location**: 61°58'N; 121°49'W. Approximately 70 m south of Mackenzie Highway kilometre post 504.7. Site A (Fig. 4a), 30 m west of old cutline in spruce forest; site R (Fig. 4b) on cutline. **Geology**: Silt of lacustrine plain. **Vegetation**: Site A in upland hardwood spruce forest, some larch and aspen, thick sphagnum ground cover. Site R, overgrown cutline, aspen 10 cm diameter, peat/mineral soil.*

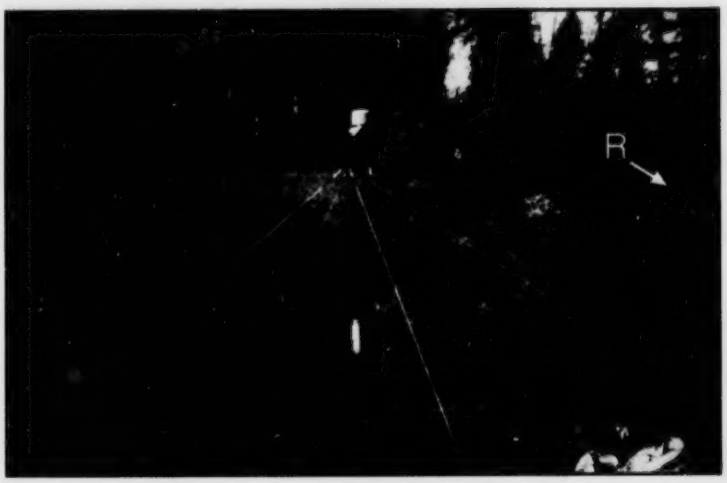

Figure 5.

Mackenzie Hwy. kp 507.6
Site A: n_{th} = 0.66 n_{fr} =0.12
Site R: n_{th} = 0.54 n_{fr} =0.45
***Location**: 61°58'N; 121°52'W. Approximately 100 m west of Mackenzie Highway kilometre post 507.6. The air screen is in small bog, whereas site R is on upland ridge beyond bog (at arrow). **Geology**: Bog developed on lacustrine plain. **Vegetation**: Area of mast, small hummocky bog, with standing pools of water in July 1993 but drier in July 1994. Site R, in mineral soil with upland spruce, juniper. Both sites open onto fen to west (to left of photo).*

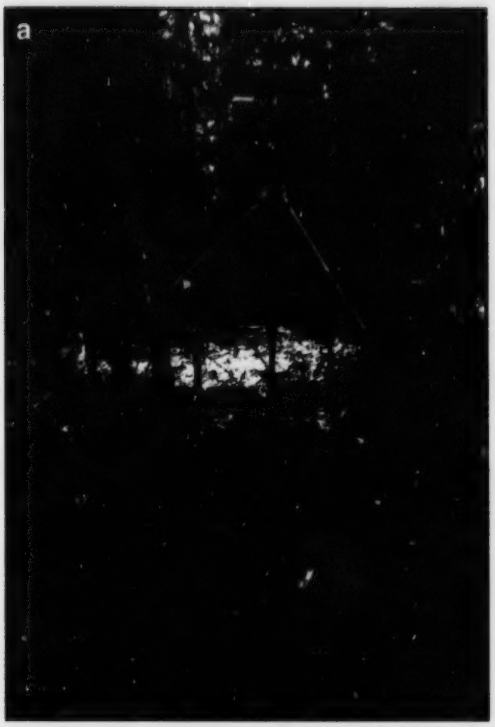

Figure 6. *River Between Two Mountains*
Site A: $n_{th} = 0.44$ $n_{fr} = 0.36$
Site R: $n_{th} = 0.26$ $n_{fr} = 0.29$
Location*: 62°57'40"N; 123°12'32"W*
Elevation*: 100 m.*
Site A (fig. 6a), top of bank of small tributary stream of Mackenzie River downstream of River Between Two Mountains. Site R (Fig. 6b), 50 m east of site A in bog bordered by fen. ***Geology****: Silt on alluvial terrace. 15–20 cm fibrous rooty organic material grading to very organic, oxidized silt. Site A, well drained.* ***Vegetation****: Site A, tall open hardwood and spruce upland forest. White spruce, birch mature forest with aspens and willows. Site R, bog.* ***Active Layer****: Site A, maximum 0.75 m (1993).*

Figure 7*. Saline River*
Site A: $n_{th} = 0.72$ $n_{fr} = 0.28$
Location*: 64°17'39"N; 124°31'07"W. North side of Saline River, 0.7 km from Mackenzie River. On approximately 30° south-facing slope above 30 m cutback.* ***Geology****: Spur cut in high stony alluvial terrace. Well drained.* ***Vegetation****: Open aspen grove. Moss, buffalo berry, northern bedstraw, bush cranberry, some juniper, alder, rose. Cover 100%.*

installation in July, 1993 but dry in July, 1994. Figure 6, River Between Two Mountains: this site illustrates how both freezing and thawing n-factors are higher in the hardwood/spruce forest (site 'A') than in the open fen some 30 m distant (site 'R'). This may result from higher evapotranspiration and a longer 'zero curtain' (Fig. 1) at the wetter fen site. Figure 7, Saline River: $n_{th} = 0.7$ on a steep (~30°) south-facing slope in open aspens; a low cranberry cover may keep the n-factor from being higher under this exposure. Figure 8, Norman Wells AES: $n_{th} = 0.72$ in the short grass of the cleared area of the AES compound (site 'R'), while about 100 m further north in bush and moss, $n_{th} = 0.34$ (site 'A').

The above relationships tend to hold in the full data set, as apparent by the clustering in a cross plot of n_{th} and n_{fr} (Fig. 9). Also, n-factors for the freezing season are lower in most cases than the corresponding thaw-season values, reflecting the reduced impact of air temperatures and solar radiation on the ground surface due to the addition of an insulating snow cover. Of 27 determinations in the full data set,

$$n_{th} \text{ (average)} = 0.5 \pm 0.2$$
$$n_{fr} \text{ (average)} = 0.3 \pm 0.1$$

Figure 8. *Norman Wells AES*
Site A: $n_{th} = 0.34$ $n_{fr} = 0.13$
Site R: $n_{th} = 0.72$ $n_{fr} = 0.13$
Location*: 65°17'35"W; 126°45'40"W*
Site A (Fig. 8a) in bush area 0.2 km north of AES Upper Atmosphere Station. Site R (fig. 8b) in AES Upper Air Station compound. ***Geology****: Till plain.* ***Vegetation****: Site A, low, open black spruce and hardwoods, low white spruce, some alder and willow. Feather moss. Ground cover of labrador tea, blueberry, bush cranberry, Saskatoon berry, rose bushes, horsetail. Site R, cleared area, short grass, about 20 cm organic mat remains.* ***Active layer****: Deeper than 1.2 m in mid-July (note frost probe handle to right of stake).*

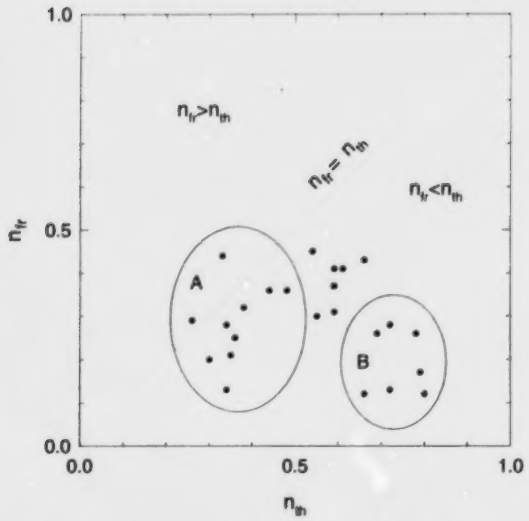

Figure 9. *Cross plot of* n_{th} *and* n_{fr}. *There is no functional relation between these two parameters, but this diagram shows the clustering that might be attributed to vegetation. Cluster A, shady bush sites, and usually with thick moss cover; cluster B, more open sites (old cutlines or open forest).*

While these values are preliminary and pertain, strictly speaking, to the 1993–1994 year and to the sites of measurement, the intent is that they might be generally representative of similar sites within the region. 'Equivalent sites' might be those having a similar site description or the same forest classification.

EXAMPLE APPLICATION

In permafrost regions, the depth of the soil that seasonally thaws (the active layer, *see* Nixon (2000)) may be estimated from air temperatures and the thermal properties of the ground if an appropriate n-factor is used to account for the effect of vegetation and snow cover in the Stefan equation (Lunardini, 1981):

$$Thaw\ depth = \lambda \sqrt{2 \frac{k_{th}}{L} I_{s,th}}$$

where k_{th} is the unfrozen soil thermal conductivity and L the latent heat of fusion of the water or ice in the soil. The term λ accounts for the need to warm the soil before thawing can commence and is a function of the soil properties and climate;

for the upper Mackenzie valley, $\lambda = 0.7–0.8$. As $I_{s,th}$ is generally not known, it can be calculated from the air-thawing index available at nearby weather stations: $I_{s,th} = n_{th} * I_{a,th}$.

As an example, we estimate the contrast in maximum active-layer thicknesses between the natural forest (site 'A') and the cleared area (site 'R') at the Norman Wells upper atmosphere station, using values of n-factors measured at these sites (Fig. 8). Estimates of other parameters needed in the equation are: $k_{th} = 2$ W/mK and $L = 6.7 \times 10^7$ J/m^3 for till with 20% volumetric water content, $\lambda = 0.7$ and $I_{a,th} = 1670$ degree-days for Norman Wells weather station (Fig. 3). For the forest site, $n_{th} = 0.34$ so $I_{s,th} = 0.34 \times 1670 = 568$ degree-days or 4.9×10^7 degree-seconds. Note that the ground surface beneath the vegetation and snow receives only one third of the degree-days of thawing potential 'available' in the air at this site. Hence,

$$\textit{Thaw depth, forest} = 0.7\sqrt{\frac{2 \times 2.0 \times 4.9 \times 10^7}{6.7 \times 10^7}} = 1.2 \text{ m}$$

$$\textit{Thaw depth, cleared site} = 0.7\sqrt{\frac{2 \times 2.010.4 \times 10^7}{6.7 \times 10^7}} = 1.7 \text{ m}$$

If there was no attenuation due to the grass at this latter site (i.e. $n_{th} = 1.0$), the maximum thaw would be deeper (2.1 m).

ACKNOWLEDGMENTS

Invaluable advice and logistic assistance in the field were provided by Pat Wood and staff, Water Resources Branch, Environment Canada, and B. Gauthier, Indian and Northern Affairs Canada, both in Fort Simpson, Northwest Territories. V. Allen designed the air temperature masts. Funding was provided by the Geological Survey of Canada under the Green Plan, and by the Panel for Energy Research and Development, Natural Resources Canada. V. Lunardini is thanked for his thorough review of the manuscript.

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Thaw-depth monitoring

F.M. Nixon[1]

Nixon, F.M., 2000: Thaw-depth monitoring; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 119–126.*

Abstract: Fifty-eight sites have been established along a 1200 km transect in the Mackenzie valley to monitor processes linking climate, climate change, permafrost, and the active layer. Annual maximum thaw penetration and surface movement are measured relative to thaw tubes anchored in permafrost. Active layer thickness, calculated from thaw penetration and surface movement, varies as a result of local soil properties, vegetation, and microclimate as well as regional atmospheric climate. Whereas thaw penetration has increased at most sites over the last 6–8 years, this increase is not always reflected by an equivalent increase in active layer thickness because of thaw settlement. Frequent air and shallow ground temperatures provide a complementary record at many sites. Active layer thickness influences biotic soil conditions, foundation performance, slope stability, and hydrology. Monitoring the active layer can contribute to modelling near surface permafrost and the impact of environmental change and can assist adaptation.

Résumé : Pour surveiller les processus faisant intervenir le climat, le changement climatique, le pergélisol et le mollisol, on a établi 58 sites le long d'un transect de 1 200 km traversant la vallée du Mackenzie. On mesure la pénétration maximale du front de dégel annuel et le déplacement du sol superficiel au moyen de tubes de dégel ancrés dans le pergélisol. L'épaisseur du mollisol, calculée à partir de la pénétration du front de dégel et du déplacement de la surface du sol, varie en fonction des propriétés locales du sol, de la végétation, du microclimat et du climat atmosphérique régional. Même si la pénétration du front de dégel a augmenté à la plupart des sites au cours des six à huit dernières années, cela ne se traduit pas toujours par un accroissement équivalent de l'épaisseur du mollisol en raison d'un tassement dû au dégel. Pour compléter ces données, on a mesuré les températures de l'air et du sol à faible profondeur à de nombreux sites. L'épaisseur du mollisol influe sur les conditions biotiques du sol, la solidité des fondations, la stabilité des versants et l'hydrologie. On peut se servir des observations faites sur le mollisol pour modéliser le pergélisol proche de la surface et les répercussions d'un changement environnemental et ainsi faciliter la prise de mesures d'adaptation.

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INTRODUCTION

One feature of permafrost that has responded significantly to past climate change is thickness of the active layer (Mackay, 1975, 1976). The active layer, overlying permafrost, is earth material that thaws from the surface and refreezes each year (Fig. 1). It forms the interface between permafrost and the atmosphere and biosphere, which host many human activities. Active-layer thickness influences vegetation and soil conditions, thereby influencing hunting, gathering, forestry, and agriculture. Thickness, texture, and moisture content of the active layer affects foundation conditions for transportation and construction. Changes in the active layer can contribute to slope instability, with impacts on transportation facilities and other structures. It is important to understand how the active layer varies locally and regionally, and how it will respond to environmental change.

LOCATION AND SITE SELECTION

Figures 2a and 2b show the location of 58 active-layer monitoring sites that have been established in the Mackenzie valley from Fort Simpson to the Arctic coast (Nixon and Taylor, 1994; Nixon et al., 1995). As the development of the active

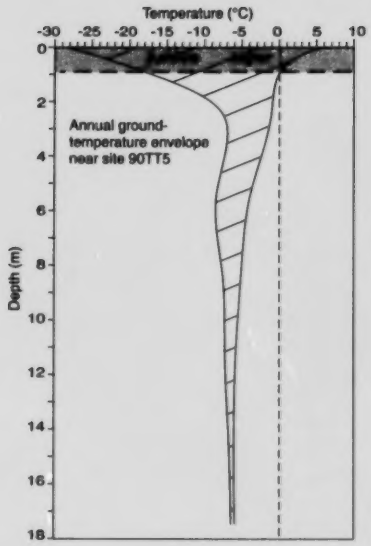
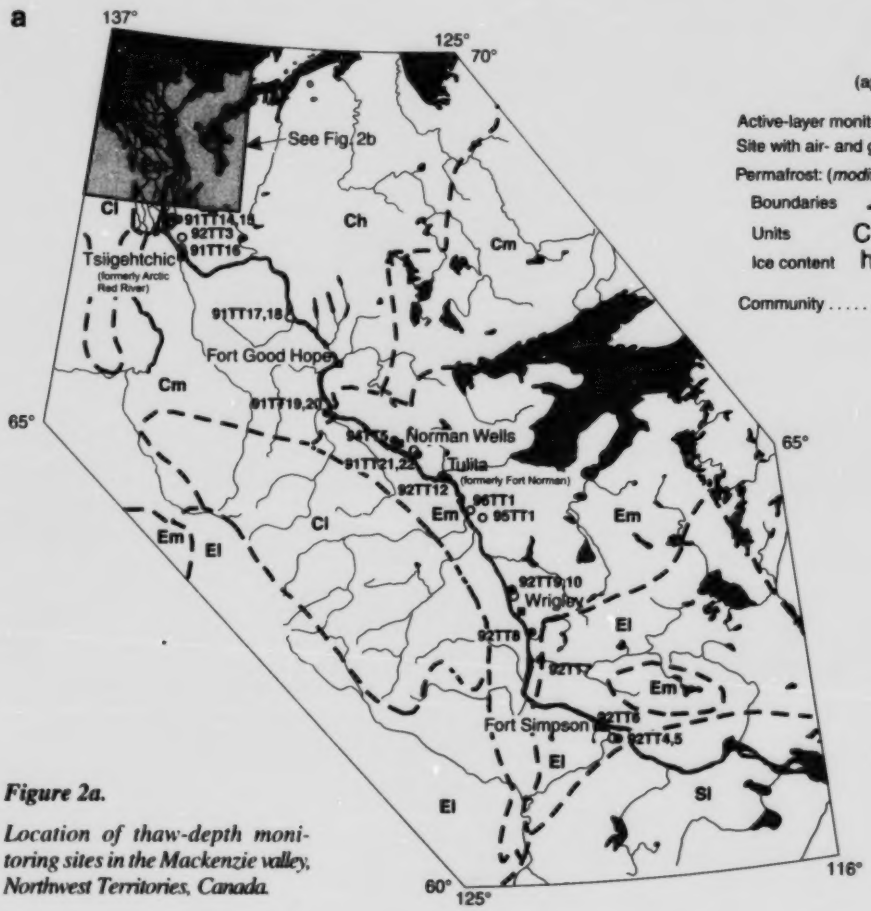


Figure 1. *Annual ground-temperature variation and resulting active layer near site 90TT5 on Richards Island, Mackenzie Delta.*



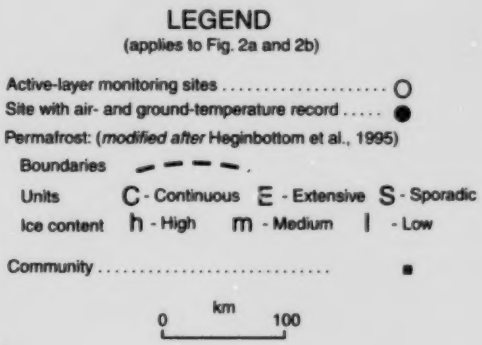

LEGEND
(applies to Fig. 2a and 2b)
Active-layer monitoring sites
Site with air- and ground-temperature record
Permafrost: (modified after Heginbottom et al., 1995)
Boundaries
Units C - Continuous E - Extensive S - Sporadic
Ice content h - High m - Medium l - Low
Community
0 km 100

Figure 2a.

Location of thaw-depth monitoring sites in the Mackenzie valley, Northwest Territories, Canada.

layer is affected by climate, vegetation, geomorphology, hydrology, and soil properties, undisturbed sites were chosen to reflect the diversity of these factors. This system will provide a sample of the variability of active-layer thickness in space and time for the Mackenzie valley.

A number of measurements and observations were made initially at each station to characterize the local environment. Landform and surface morphology were described, slope and aspect were estimated, and elevation measured. A description of vegetation was made and soil sampling was carried out to or below the frost table for laboratory determination of moisture content and texture. Some of these observations have been summarized for each station in Table 1.

Site selection and instrumentation were co-ordinated with several other studies in the region (e.g. Gibson et al, 1993; Tarnocai et al., 1995; Mackay, 1997; Quinton and Marsh, 1998; Burgess and Smith, 2000). In particular, air- and ground-temperature data gathered by Tarnocai et al. (1995) and Taylor (1995) provide continuous thermal records at many thaw-tube sites. The transect crosses several ecoclimatic regions (Ecoregions Stratification Working Group, 1989) as well as permafrost boundaries (Heginbottom et al., 1995) and the sites have been selected to provide maximum variety of natural situations while remaining accessible within the Mackenzie valley transportation corridor.

INSTRUMENTATION

Maximum annual thaw penetration and maximum heave and subsidence of the ground surface are measured using a modified version of a frost tube developed by Mackay (1973). The device is a water-filled observation tube located in a heave-resistant access tube (Fig. 3). The ice-water interface in the observation tube corresponds to the frost table in the surrounding ground. A marker (glass bead) rests on the ice surface, descending during the thaw season to be trapped at maximum depth on freeze back in late summer or fall. Through the use of very light equipment and access to sites on foot, disturbance to the vegetation and soils during installation and observation is minimized. Most sites show nearly complete recovery after one or two seasons (Fig. 4).

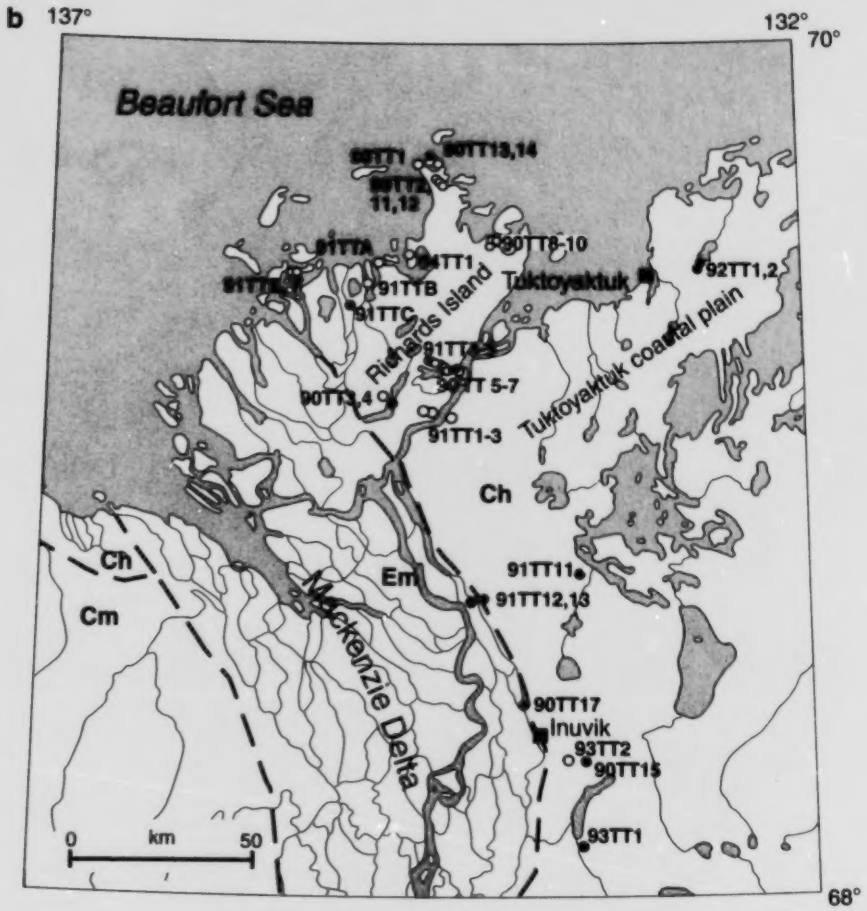


Figure 2b.

Locations of thaw-depth monitoring sites in the Mackenzie Delta area.

Table 1. Site location and active layer depths, 1993 to 1998 thaw seasons in order of increasing latitude.

Site	Latitude/ longitude	Elevation (m)	Active layer (cm)	Environmental conditions
92TT4	61°46′13″ 121°11′33″	126	101, 109, 118, 118, 118, 121	Open hardwood forest, 60% ground cover, 30 cm organic layer, well drained, riverbank (river ≈120 m a.s.l.), SW aspect, 85 cm snowpack
92TT5	61°45′58″ 121°11′6″	142	160, 174, 177, 182, 183, 189	Aspen thicket, 10 cm organic layer, well drained, inland, 55 cm snowpack
92TT6	61°53′13″ 121°35′59″	≈165	95, 106, 123, >128, n.a., n.a.	Open spruce forest, complete ground cover, 25 cm organic layer, imperfect drainage, inland
92TT7	62°41′48″ 123°3′46″	103	79, 83, 84, 90, 87, 89	Open spruce with hardwood forest, 100% ground cover, 25 cm organic layer, imperfect drainage, river bar inland from Mackenzie River (Willowlake River ≈95 m)
92TT8	62°57′40″ 123°12′32″	99	75, 76, <80, 81, >89	Open hardwood with spruce forest, complete ground cover, 20 cm organic layer, well drained, bank 6 m above river, southern aspect
92TT9	63°27′40″ 123°41′42″	97	<52, 52, 52, 51, 49, 49	Open spruce with hardwoods, complete ground cover, 40 cm organic layer, poorly drained, bank 7 m above river, 35 cm snowpack
92TT10	63°27′59″ 123°41′35″	97	<58, 60, 58, <64, <65, 66	Open spruce forest, complete ground cover, 30 cm organic layer, imperfect drainage, inland, 40 cm snowpack
92TT11	64°17′20″ 124°31′28″	99	96, destroyed 1994	Open spruce with hardwoods, 20 cm organic layer, well drained, steep bank 12 m above river, western aspect, 35 cm snowpack
92TT12	64°54′40″ 125°34′32″	93	<55, n.a., n.a., n.a., 61, 66	Open black spruce, complete ground cover, >45 cm organic layer, imperfect drainage, inland, gentle westerly slope, 50 cm snowpack
91TT21	65°11′29″ 126°27′35″	54	149, 149, 149, 165, >175, n.a.	Open spruce with hardwoods, complete ground cover, 15 cm organic layer, river bank, 60 cm snowpack
91TT22	65°11′52″ 126°27′56″	51	63, 61, 61, 59, 59, 66	Open spruce forest, complete ground cover, 30 cm organic layer, 65 cm snowpack
94TT5	65°17′54″ 126°50′43″	65	n.a., n.a., 96, 95, 103, 103	Open spruce with hardwoods, complete ground cover, 30 cm organic layer, fair drainage, inland
91TT19	65°39′41″ 128°46′10″	3	71, 73, 73, 71, 72, 74	Very open spruce, 98% ground cover, >35 cm organic layer, riverbank, 55 cm snowpack
91TT20	65°40′29″ 128°49′37″	39	59, 62, 62, 58, 59, 62	Hardwood thicket, complete ground cover, 15 cm organic layer, inland, base of north-facing scarp, 60 cm snowpack
91TT17	66°47′16″ 130°07′15″	22	85, n.a., 86, (n.a. after 1995)	Open spruce, no organic layer, well drained, riverbank, 50 cm snowpack
91TT18	66°47′10″ 130°07′6″	24	75, 78, 80, n.a., n.a., 91	Scattered black spruce, complete ground cover, 15 cm organic layer, imperfect drainage, inland, 50 cm snowpack
91TT16	67°29′03″ 133°45′47″	14	93, 97, 98, 85, 88, 96	Open high shrub, 75% ground cover, no organic layer, well drained, river bar, 65 cm snowpack
92TT3	67°29′44″ 133°45′47″	≈30	n.a., 73, 75, 76, 76, 83	Open black spruce, complete ground cover, 10 cm organic layer, imperfect drainage, 65 cm snowpack
91TT14	67°47′42″ 134°07′34″	13	106,111, 111, 108, 110, 116	Open spruce with hardwoods, complete ground cover, no organic layer, well drained, riverbank
91TT15	67°45′58″ 134°04′43″	8	75, 80, 76, 74, 74, 80	Open spruce with hardwoods, 95% ground cover, no organic layer, imperfect drainage, inland lakeshore
93TT1	68°06′39″ 133°28′29″	≈30	68, 54, 58, 57, 64, 69	Scattered black spruce, complete ground cover, inland, 40 cm snowpack
90TT15	68°18′57″ 133°25′51″	83	74, 90, 88, 87, 86, 101	Open bog, organics 2.5 m thick, inland, 75 cm snowpack
93TT2	68°19′13″ 133°30′48″	≈80	n.a., 55, 59, 60, 58, 65	Scattered black spruce, complete ground cover, 12 cm organics, imperfect drainage, inland, 70 cm snowpack
90TT17	68°24′59″ 133°47′13″	5	n.a., 128, >138, (n.a. after 1995)	Open high shrub, incomplete ground cover, no organic layer, well drained, delta-surface channel bank, 60 cm snowpack
91TT11	68°44′13″ 133°29′15″	53	69, 68, 68, 67, <68, 69	Shrub tundra, complete ground cover, 10 cm organic layer, inland, 40 cm snowpack
91TT12	68°41′24″ 134°06′18″	152	70, 76, 71, 71, 71, 73	Low-shrub tundra, partial ground cover with bare hummock surfaces, no organic layer, inland, 13 cm snowpack
91TT13	68°41′03″ 134°08′20″	5	129, 132, 136, 134, 135, 140	High shrub, incomplete ground cover, no organic layer, well drained, delta-surface bar, 60 cm snowpack

The active layer is defined as the thaw recorded in the thaw tube, minus the height of the tube above the ground surface at maximum surface subsidence, assumed to occur about the time of maximum thaw. Minimum values indicate that the value of maximum subsidence was not available, and the ground-surface height measured during mid-summer in the year following the record is used as a minimum subsidence at time of maximum thaw. Sites are arranged from south to north. Snowpack observations were taken during March or April, and snowpack is reported as a typical value from all the observations at the site. n.a. — not available.

Table 1 (cont.)

Site	Latitude/ longitude	Elevation (m)	Active layer (cm)	Environmental conditions
90TT3	69°08′44″ 134°42′45″	40	96, 102, 101, 100, <102, 104	Low-shrub tundra, 80% ground cover, no organic layer, well drained, inland hilltop, 10 cm snowpack
90TT4	69°08′44″ 134°41′53″	10	87, 93, 91, 92, <95, 97	Shrub tundra, complete ground cover, no organic layer, near lakeshore, 25 cm snowpack
91TT1	69°06′30″ 134°23′56″	8	61, 64, 64, 59, <66, 67	Shrub tundra, complete ground cover, 10 cm organic layer, imperfect drainage, inland, gentle east-facing slope, 38 cm snowpack
91TT2	69°06′35″ 134°22′50″	3	75, 76, 77, 76, <80, 83	High shrub, incomplete ground cover, no organic layer, imperfect drainage, floodplain, 95 cm snowpack
91TT3	69°06′31″ 134°20′53″	11	46, 50, 48, 46, <49, 49	Shrub tundra, complete ground cover, 30 cm organic layer, well drained, west-facing slope, 29 cm snowpack
91TT5	69°14′27″ 134°25′52″	53	78, 87, 89, 83, <94, 98	Low-shrub tundra, disturbed ground cover, cryoturbated organics to 25 cm, inland hilltop, 10 cm snowpack
91TT6	69°14′28″ 134°25′09″	23	46, 48, 46, 46, <52, 52	Low-shrub polygonized wetland, 75% ground cover, >20 cm organic layer, inland basin, 5 cm snowpack
91TT7	69°14′26″ 134°24′13″	37	55, 62, 64, 64, <70, 78	Low-shrub tundra, 80% ground cover, >20 cm cryoturbated organics, imperfect drainage, inland hilltop, 10 cm snowpack
91TT4	69°13′45″ 134°20′58″	25	110 , ('91 only)	Grass sedge wetland, complete ground cover, no organic layer, recently drained lake basin, inland, 25 cm snowpack
91TT8	69°13′35″ 134°20′12″	≈50	61, 56, 61, 57, <63	Low-shrub tundra, inland hilltop, 20 cm snowpack
91TT9	69°12′58″ 134°17′40″	16	41, 40, 38, 38, <39, 84	Shrub-sedge polygonized wetlands, complete ground cover, inland basin, 20 cm snowpack
90TT5	69°13′06″ 134°17′05″	39	80, 86, 85, 75, <89, 91	Low-shrub tundra, complete ground cover, 5 cm organic layer, inland hilltop, 20 cm snowpack
90TT6	69°13′01″ 134°16′40″	9	58, 61, 60, 46, <61, 64	Shrub tundra, complete ground cover, >25 cm organics, riverbank, 30 cm snowpack
90TT7	69°13′04″ 134°16′25″	1	77, 78, 75, 61, n.a., 75	Sedge wetland, complete ground cover, no organic layer, saturated, water covered, floodplain, 20 cm snowpack
92TT1	69°28′22″ 132°37′37″	25	48, 47, 48, 47, 48, 48	Shrub tundra, complete ground cover, 20 cm organic layer, well drained, inland hilltop, 20 cm snowpack
92TT2	69°28′27″ 132°38′04″	4	<44, 49, 50, <53, 53, 60	Shrub wetland, complete ground cover, 10 cm organic layer, inland thermokarst basin, 53 cm snowpack
90TT8	69°31′48″ 134°00′57″	14	74, 76, 77, <70, <83, 87	Low-shrub tundra, 75–80% ground cover, no organic layer, north-facing slope, 5 cm snowpack
90TT9	69°31′51″ 134°00′21″	4	72, 70, 70, 64, <79, 91	Low-shrub tundra, complete ground cover, 90 cm organic layer, coastal, 5 cm snowpack
90TT10	69°31′42″ 134°02′15″		76, 86, 84, <77, <90, 90	Low-shrub tundra, 75–80% ground cover, no organic layer, south-facing slope, inland, 10 cm snowpack
90TT2	69°39′46″ 134°23′07″	29	54, 53, 51, 50, <53, 56	Low-shrub tundra, 95% ground cover, organic layer to 45 cm, inland hilltop
90TT11	69°39′35″ 134°22′34″	16	63, 65, 64, 58, n.a., 70	Low-shrub tundra, complete ground cover, organics from 30 to 45 cm, moderate east slope, inland
90TT12	69°39′25″ 134°21′44″	3	59, 57, 59, 53, <57, 56	Low-shrub tundra wetland, 80% ground cover, mixed organics and silt to 80 cm, coastal
94TT1	69°29′07″ 134°33′43″	5	n.a., n.a., 63, 61	Low-shrub tundra, complete ground cover, mixed silt and organics to 110 cm, inland, 20 cm snowpack
90TT1	69°42′53″ 134°29′02″	17	<56, 53, n.a., <49 (n.a. after 1996)	Low-shrub tundra, complete ground cover, 35 cm organic layer, high coastal bluff
90TT13	69°43′18″ 134°27′14″	≈3	63, 64, 62, 60, <66, 70	Low-shrub tundra, complete ground cover, mixed silt and organics to 40 cm, low coastal, 13 cm snowpack
90TT14	69°43′07″ 134°26′29″	15	<81, 82, 79, (n.a. after 1995)	Low-shrub tundra, 75% ground cover, mixed silt and organics to 35 cm, well drained, inland scarp edge, 5 cm snowpack
91TTA	69°28′53″ 134°49′43″	< 2	75, 78, 77, 70, n.a., 77	Sedge, grass wetland with low shrub, 100% ground cover, very organic silt 30 cm, outer delta, coastal, 34 cm snowpack
91TTB	69°24′36″ 134°50′35″	< 2	117, 113, <123, 104, (n.a. after 1996)	Open high shrub, 60% ground cover, no organic layer, outer delta, inland
91TTC	69°22′09″ 134°56′55″	< 2	<119, 118, >124, <112, 118, n.a.	High shrub, 50% ground cover, no organic layer, well drained, outer delta, inland, 70 cm snowpack
91TTE	69°26′34″ 135°21′53″	< 2	n.a., < 99, 91, (n.a. after 1995)	Grass and sedge wetland, no organic layer, outer delta island, coastal
91TTF	69°26′55″ 135°20′22″	< 2	112, 117, 114, 95, 107, 150	Grass and sedge wetland, 75% ground cover, no organic layer, saturated, under shallow water, outer delta island, coastal

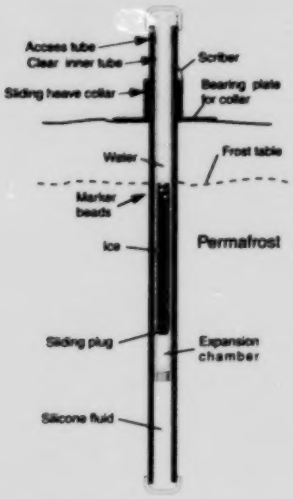


Figure 3. *Apparatus used to record maximum thaw penetration and maximum ground-surface movement at active-layer monitoring sites. Drawing is not to scale: length is about 4 m, diameter of outer tube is about 3 cm.*

Figure 4. *Thaw-depth monitoring site (92TT8) one year after installation, showing minimal disturbance of ground surface. Photograph by F.M. Nixon. GSC 2000-032*

RESULTS AND DISCUSSION

The active-layer thickness, reported in Table 1, is calculated by subtracting the distance between the tube top and the ground surface at maximum subsidence, recorded by the heave/subsidence scriber, from the distance between the tube top and last season's marker bead. This assumes maximum penetration of thaw corresponds to maximum ground-surface subsidence.

During the summer of 1994, the first complete survey of the active-layer monitoring system in the Mackenzie valley was achieved. A preliminary analysis of the data indicates that active layers, ranging from 38 cm to 182 cm, vary from site to site as a result of a complex interplay of site-specific and regional factors. The intuitive correspondence between active-layer thickness and latitude (Fig. 5) is statistically significant, though weak (correlation coefficient = 0.44). This test was applied after the removal of Mackenzie Delta and recently drained tundra lake sites showing anomalously thick active layers due to locally warmer ground-temperature profiles (Smith, 1976; Mackay, 1985). The scatter may be, in part, due to the fact that thawing air temperatures do not vary continuously with latitude at undisturbed temperature monitoring sites in the Mackenzie valley (Tarnocai et al., 1995). The influence of local factors, such as organic cover and snow depth, will also mask the effect of latitudinal change in climate (Burn, 1998). When tested, neither snow nor organic cover showed a significant correlation with thickness of active layer. Despite this, at some adjacent sites, for instance 90TT5/90TT6 and 91TT1/91TT3, it is apparent that organic cover is the most effective local factor controlling active-layer development (Table 1). Other local factors such as snow depth and density are certainly just as dominant elsewhere.

Data are available from as early as 1990 to 1996 for selected sites, and these can be used to display interannual variation of thaw at each (Fig. 6). The parameter plotted on this graph is not active-layer thickness, but maximum thaw penetration measured from a standard reference point that is assumed stable and independent of the moving ground surface. Records that may contain a component of heave have been discarded. The choice of this parameter for comparison of thaw from year to year simplifies the interpretation by eliminating such complexities as ground-ice melting and soil compaction. Most sites show a steady increase in maximum thaw until 1996, a trend also reported for soil climate sites throughout the valley (Tarnocai et al., 1995). This may reflect warmer air temperatures in the area. Summer average air temperatures between 1991 and 1992, recorded at three automatic weather stations on Richards Island, increased consistently by 2°C. (A.S. Judge, unpub. data, 1994). A longer 1993 thaw season, higher annual average temperature (+0.6°C), and higher summer average (+1.8°C) is reported for the Arctic Tundra Canadian climate region in which these sites are located. The adjacent Mackenzie climate region also experienced a longer thaw season and higher annual average temperature in 1993 (Fig. 5, 7, 8 *in* Skinner and Maxwell, 1994). Summer averages for tundra soil climate sites where higher in 1994 than any year since 1990, and shallow summer soil temperatures at

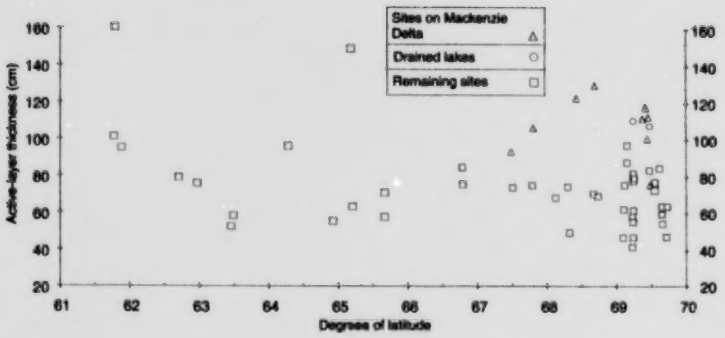


Figure 5.

Plot of thaw depth versus latitude for each monitoring site in 1993.

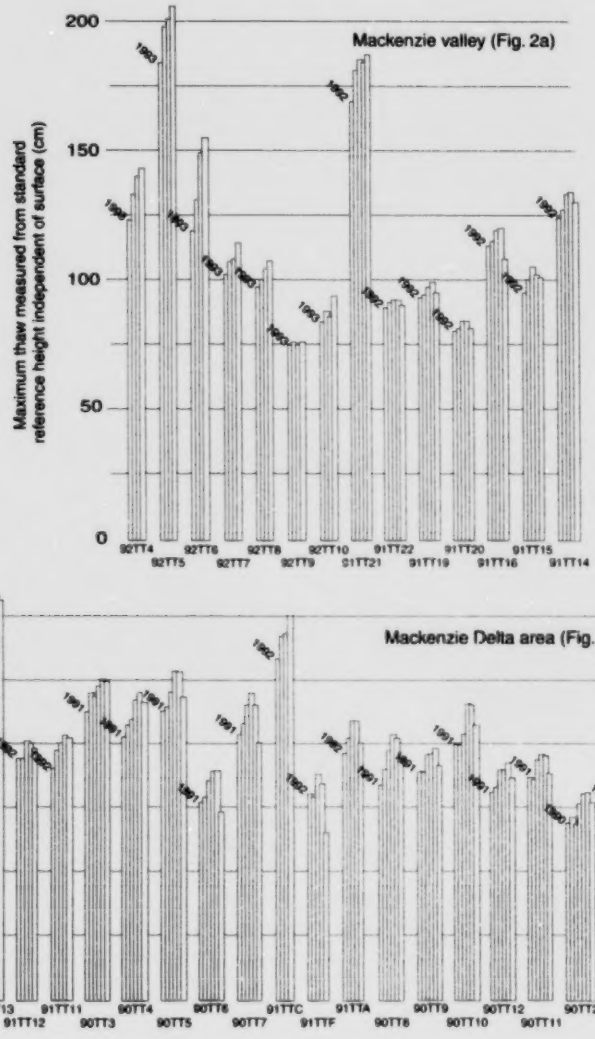

Mackenzie valley (Fig. 2a)
Maximum thaw measured from standard reference height independent of surface (cm)
Mackenzie Delta area (Fig. 2b)
Maximum thaw measured from standard reference height independent of surface (cm)

Figure 6.

Year-to-year variability of the maximum thaw for selected sites up to 1996, grouped according to Figure 2a and 2b. The year record began is indicated.

the same sites were as warm as any year on record (Tarnocai et al., 1995). A decrease in thaw penetration in 1996 at sites north of Norman Wells corresponds to the coldest summer mean temperature for this area in more than four years, with August and September being about 2°C cooler than normal (Canadian Meteorological Centre, 1996). The record from thaw-depth monitoring tubes is presently too short to make any inference about temporal trends at a climate scale.

Changes to active-layer thickness will influence surface stability through thaw settlement, frost heave potential, and bearing capacity. Slope stability will also be affected (Aylsworth and Egginton, 1994). Besides these geotechnical characteristics, changes in active-layer thickness will affect hydrology (Hinzman and Kane, 1992) and, through a combination of enhanced mass-movement disturbance, changes in soil moisture (Edlund et al., 1989), and nutrient availability (Waelbroeck et al., 1997), can affect the ecology of an area through modified vegetation. Monitoring of annual active-layer development in a variety of settings and analysis of how this development relates to widely available or easily measured parameters will provide field data to test models which include active-layer response in their portrayal of climate-change effects.

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Permafrost and surficial materials along a north-south transect: observations from the Norman Wells pipeline

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Burgess, M.M. and Lawrence, D.E., 2000: Permafrost and surficial materials along a north-south transect: observations from the Norman Wells pipeline; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 127–141.*

Abstract: The Norman Wells pipeline, the first completely buried oil pipeline in permafrost in Canada, traverses 869 km of the discontinuous permafrost zone from Norman Wells, Northwest Territories to Zama, Alberta. Observations of permafrost and terrain conditions were recorded during the excavation of the 1.2 m deep pipeline trench. These records provide a unique detailed transect of near-surface permafrost extent, ice content, and sediment characteristics over terrain both newly cleared for the pipeline project as well as previously cleared right-of-ways. Frozen to unfrozen transitions average about two per kilometre, increasing to four or five transitions per kilometre in southern peatlands. Frozen ground distribution ranges from about 80% of right-of-way length at the north end to about 30% at the south end, although locally there is wide variation. Visible ice is present in essentially all permafrost encountered, averaging about 20% by volume but commonly up to 40%.

Résumé : L'oléoduc de Norman Wells, le premier oléoduc enfoui dans le pergélisol au Canada, traverse une zone de pergélisol discontinue sur une distance de 869 km entre Norman Wells (T.N.-O.) et Zama (Alberta). Pendant l'excavation de la tranchée de 1,2 de profondeur pour enfouir l'oléoduc, on a consigné les observations faites sur le pergélisol et les conditions du terrain. Ces observations ont permis d'établir un profil détaillé exceptionnel de l'étendue du pergélisol proche de la surface, de la teneur en glace et des caractéristiques des sédiments dans les terrains qui ont été dégagés récemment pour le projet de l'oléoduc et dans d'autres qui l'ont été plus anciennement pour la construction des emprises de l'oléoduc. On distingue en moyenne deux zones de transition gelées à non gelées par kilomètre; ce nombre passe à 4 ou 5 par kilomètre dans les tourbières méridionales. La proportion de gélisol varie d'environ 80 % dans la partie nord du tracé à environ 30 % dans la partie sud, mais on observe, par endroits, des variations plus grandes. De la glace est visible dans tous les pergélisols traversés, y occupant en moyenne 20 % du volume; il est fréquent qu'elle occupe jusqu'à 40 % du volume.

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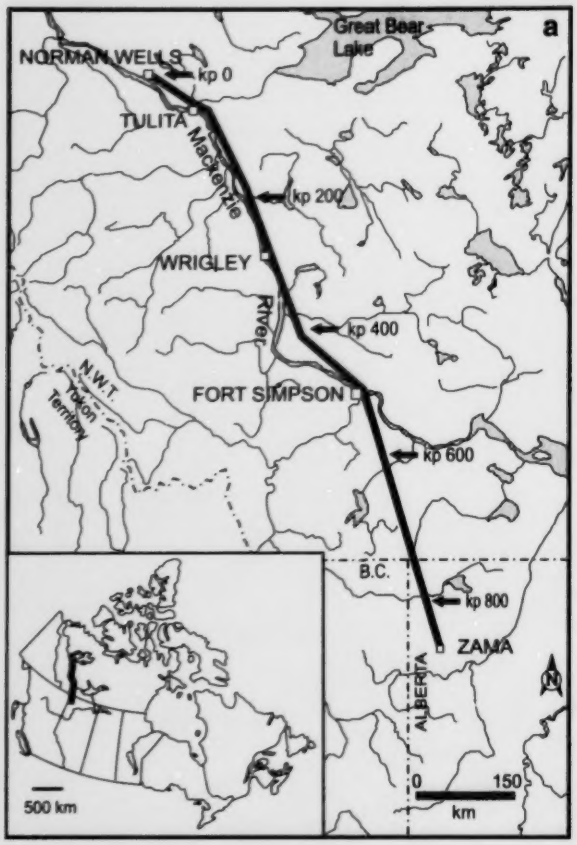
INTRODUCTION

The Norman Wells pipeline, the first completely buried oil pipeline in permafrost in Canada, is owned and operated by Enbridge Pipelines (NW) Inc. (Enbridge; formerly Interprovincial Pipe Line (IPL)) and has been in operation since April 1985. The pipeline route provides a unique and detailed north-south transect through the discontinuous permafrost zone in the Mackenzie valley (Fig. 1). Permafrost varies from widespread in the northern portion of the route, to sporadic in the south (*see* Heginbottom, 2000). The underlying sequences of unconsolidated Quaternary deposits are often ice rich and thaw sensitive (*see* Aylsworth et al., 2000). Preconstruction geotechnical investigations and design studies included a continuous geophysical survey along the right-of-way (ROW), mapping of surficial geology at a scale of 1:20 000, thaw-depth and settlement surveys and analyses, drilling of geotechnical boreholes, and geothermal modelling. A summary of the environmental and engineering considerations and a discussion of the novel design features for the pipeline can be found in Nixon et al. (1984), Hanna and McRoberts (1988), and MacInnes et al. (1989).

Research on the thermal regime along the pipeline and geomorphic response of the terrain to construction disturbance was reported in Pilon et al. (1989), Burgess and Harry (1990), MacInnes et al. (1990), Burgess and Riseborough (1990), Burgess (1992), and Burgess et al. (1993, 1995b). Data gathered during all phases of the project are accessible and ultimately archived (Burgess and Naufal, 1991; Burgess, 1993; Burgess et al., 1995a).

A valuable source of data is the observations of the pipeline-trench conditions made during construction, collectively known as the ditchwall logs or the ditchwall database. This database is unique in providing a detailed continuous transect through the discontinuous permafrost zone of terrain and permafrost conditions in the top 1.2 m of the ground.

The database provides base-line geoscience information for the assessment of the impact of climate change on the physical environment. The ditchwall data, along with



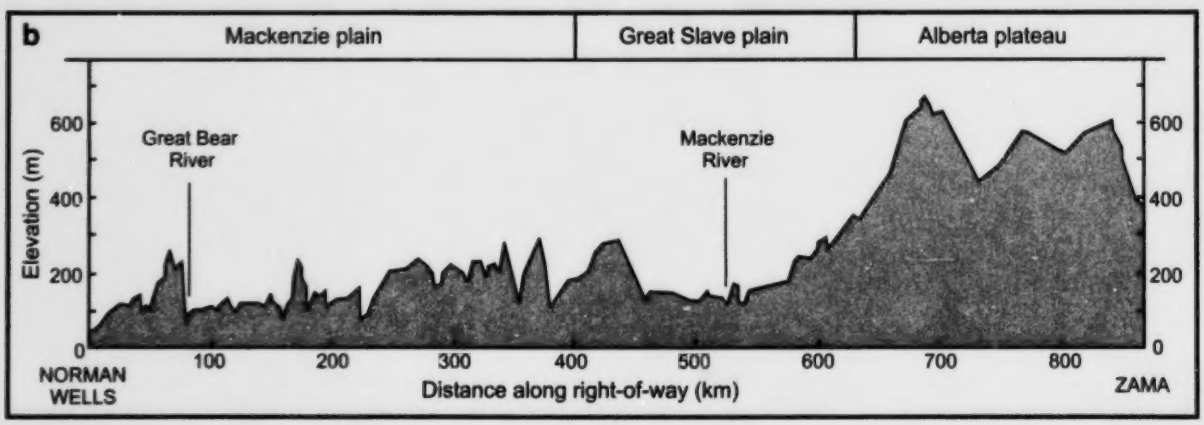


Figure 1. *a) Map showing the location of the Norman Wells pipeline route, and* ***b****) the elevation profile and physiographic regions along the route. Note: kilometrepost 0 (kp 0) is the start of the pipeline at Norman Wells, Northwest Territories and kp 869 is the southern terminus at Zama, Alberta.*

supporting information from pipeline monitoring, form the basis for the following analysis of permafrost and terrain conditions.

DITCHWALL DATABASE

Trenching for the pipeline was undertaken in the months of January, February, and March, when the active layer is refrozen or seasonal frost penetration is thick enough to allow a stable working platform for movement and operation of construction equipment. The trench depth is on average 1.2 m, and the width is generally 1 m. Depending on the segment under construction (there were six) and the nature of the terrain, various trenching methods were used: arctic ditcher (Fig. 2), conventional southern ditcher, and bulldozers and backhoes. During construction of the pipeline, conditions in the open trench were recorded by the pipeline company's inspection personnel at 100 m intervals and at major soil or permafrost interfaces. The observations in the 'ditchwall logs', recorded at 9000 locations, were used to confirm or revise site-specific design or remedial measures.

The ditchwall logs include the following basic geotechnical and permafrost information: thickness of peat, depth of trench, presence of cobbles, frozen/unfrozen condition of trench, and comments regarding soil type and ice conditions. (Note: each record, i.e. location, does not necessarily contain information in all fields.) Figure 3 illustrates soil and permafrost conditions as observed in the trench. The Geological Survey of Canada (GSC) undertook to compile the ditchwall logs into a digital database (Geo-Engineering, 1992a, b, c). Additional relevant information was added to the database, including as-built kilometrepost (kp), location (in UTM co-ordinates), terrain type (based on Enbridge's detailed alignment work (Kay et al., 1983)), and type of right-of-way clearing (previously cleared or newly cleared).

Results and highlights of the ditchwall data analyses are presented below. Some aspects of the ditchwall data have been previously examined by Nixon et al. (1991). General conditions for the whole pipeline route have been summarized using increments of 50 km: i.e. data are presented either as an average or percentage of what was encountered over each 50 km section (kp 0 is located at Norman Wells; *see* Fig. 1 for additional kp locations). A more detailed examination of several components of the database, such as frozen-ground distribution, was made at increments of 5 km, and select examples of these results are also presented. Certain data are presented on the basis of terrain unit or as a function of longer segments of the right-of-way.

TERRAIN CONDITIONS

Terrain units

The terrain classification employed for the Norman Wells pipeline is outlined in MacInnes et al. (1989). The terrain codes were simplified for entry into the digital database, and an initial examination revealed that seven units constituted approximately 85% of the pipeline route. These seven, plus an eighth category encompassing all others, formed the basis for the 50 km interval compilations, and are as follows: Lp — lacustrine landforms, generally plains, including veneers (alluvial, colluvial, eolian, and glaciofluvial) over lacustrine plain units; Mg — moraine, generally ground moraine, including veneers (alluvial, eolian, glaciofluvial) over moraine units; Lp-Mg — composite/complex of lacustrine and moraine units, also includes veneer of one over the other; Ov-Lp — organic veneer over lacustrine units; Ov-Mg — organic veneer over moraine units; Ou-Mg — composite of organic and moraine units; Ou — organic units: fen, bogs, and fen-bog complexes; Other — alluvial, colluvial, eolian, bedrock, and glaciofluvial units.

Figure 2. *Winter construction using an Arctic ditcher for trenching operations (January 1984, kp 0.2). Photograph by K.L. MacInnis. GSC 2000-033A*

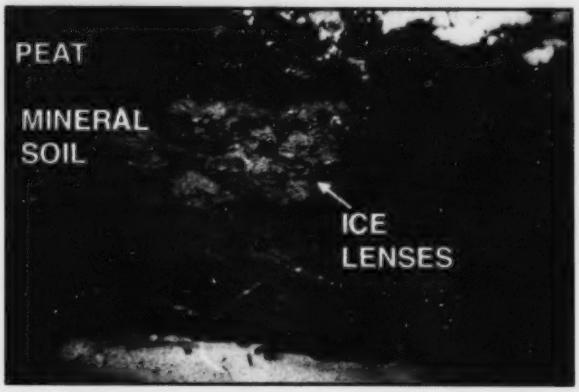


Figure 3. *Soil and permafrost conditions as observed in the trench during construction at kp 0.2. Note the thin peat layer over mineral soil and the ice lenses within the mineral soil. Photograph by D.G. Harry. GSC 2000-033B*

Figure 4a summarizes the percentage occurrence of the dominant terrain units by 50 km intervals. Lacustrine and moraine types dominate in the north (kp 0–550). Organic units or organic veneers dominate from kp 550 to 869 (*see* Aylsworth et al. (2000) for a discussion and maps of the surficial geology of the Mackenzie valley). Figure 4b provides a detailed look by 5 km increments at the first 100 km of the pipeline. At the 5 km interval, the ground is frequently dominated by an individual terrain unit reflecting the local surface geology. For example the high percentage of 'other' between kp 60 and 70 is largely bedrock as the pipeline rises in elevation and passes over Bear Rock, north of the Great Bear River. The 50 km interval represents a composite of the terrain units.

Soil groups

Soil type was recorded for all the pipeline route, except the southern section from kp 696 to 869. The thickness of the peat layer was, however, recorded in this southern section. Where peat extends to the bottom of the pipeline trench, the soil type is recorded as organic. Otherwise, soil descriptions entered in the database are, whenever possible, for the mineral soil at the bottom of the trench. Since there were eight ditchwall inspectors (with varying experience in soils identification) soil classifications were not necessarily consistent; for example the use of various terms, i.e. 'till' or 'silt-clay' may be used more or less often depending on the individual inspector or the terrain being described.

Figure 5a presents soil type summaries by 50 km interval; records with no description of soil type at the bottom of the trench appear as 'blanks'. Figure 5b links this data to the terrain-unit classification based on airphoto interpretation. The correlation is good with, for example, 65% of the Lp unit composed of clay, silt, and sand, and 60% of the Mg unit described as till.

Peat thickness

Peat thickness is recorded in most ditchwall logs, except for the section from kp 350 to 550, where only 20–30% of the records contain peat data (Fig. 6a). The thickness may be slightly underestimated, due to scraping of the surface by construction equipment to obtain a stable working platform. Figure 6a presents the average peat thickness per 50 km interval along the entire route. The plot reveals that peat thickness increases from an average of 0.3 m in the first 400 km (where lacustrine and moraine units are predominant) to an average of 0.7 in the last 450 km (where organic units/organic veneers are major units). This pattern is consistent with the distribution of organic-rich terrain units (Fig. 4a). Figure 6b provides a more detailed examination of peat thickness as a function of terrain type for three sections of the pipeline route (kp 0–375, 375–675, and 675–869).

It should be noted that in cases where peat extended to the bottom of the trench, the reported thickness may be less than the total thickness of peat, which would lead to an underestimate of the average. This can be of particular importance when examining organic terrain units. The number of observations of peat extending to the bottom of the trench was analyzed by terrain unit for each of the three pipeline route sections. For most nonorganic units, 10% or fewer of the records show peat to the bottom of the trench. In organic terrain, these minimum values for peat thickness represent 30% or more of the Ou and Ou-Mg records (and up to 65% of Ou in the kp 375–675 section), and between 10% and 25% of the Ou-Lp and Ov-Mg records.

The ditchwall database does not provide an indication of the nature of the peatlands. For a discussion of peatlands in the Mackenzie valley, *see* Aylsworth and Kettles (2000), and for a further discussion of peatlands along the pipeline route and their response to pipeline-related environmental change, *see* Burgess and Tarnocai (1997). Boreholes drilled at GSC study sites along the pipeline revealed that the thickness of peat in the frozen peat plateau bogs in northern Alberta can reach up to 7 m.

Cobble occurrence

The occurrence of cobbles and boulders in the trench base was considered a potentially serious problem that could lead to point loading of the pipe when or if settlement occurred. To alleviate this problem, select bedding or sand bags were placed under the pipeline in these areas. The criteria set to flag the presence of cobbles was the occurrence of two or more cobbles, 150 mm or greater in diameter, in 10 m of ditch. The condition was assumed to continue until no such cobbles were present for 50 m of ditch. There is some uncertainty as to how well this recording criteria was adhered to in some sections of the route. In areas where select bedding would have been used in any case (e.g. because native backfill would have been too ice rich) detailed records may not have been kept.

Figures 7a and 7b summarize cobble occurrence by 50 km intervals and by terrain type. The greatest incidence of cobbles is recorded in Lp-Mg, Mg, and Lp units, and in the northern and central portions of the pipeline route where these terrain units predominate; however, cobble occurrence recorded in some segments of the route, for example, from kp 350–450 where the Mg unit covers 50% or more of the terrain (Fig. 4a), seem low. This may reflect either a finer grained moraine in that area, or the possible discrepancies in recording as mentioned above. The high incidence in the northernmost segments may also reflect the greater concern for thaw settlement in the northern section of the route (where the percentage of terrain underlain by permafrost is greater; *see* discussion below).

PERMAFROST

Thermal conditions

The thermal condition of the trench was recorded in the following fashion: F, frozen, either the whole trench or at least the trench bottom; U, unfrozen, either the whole trench or at least the trench bottom; F/U, frozen over unfrozen; normal seasonal freezeback in unfrozen terrain or terrain with an active layer greater than trench bottom; F/U/F, frozen over

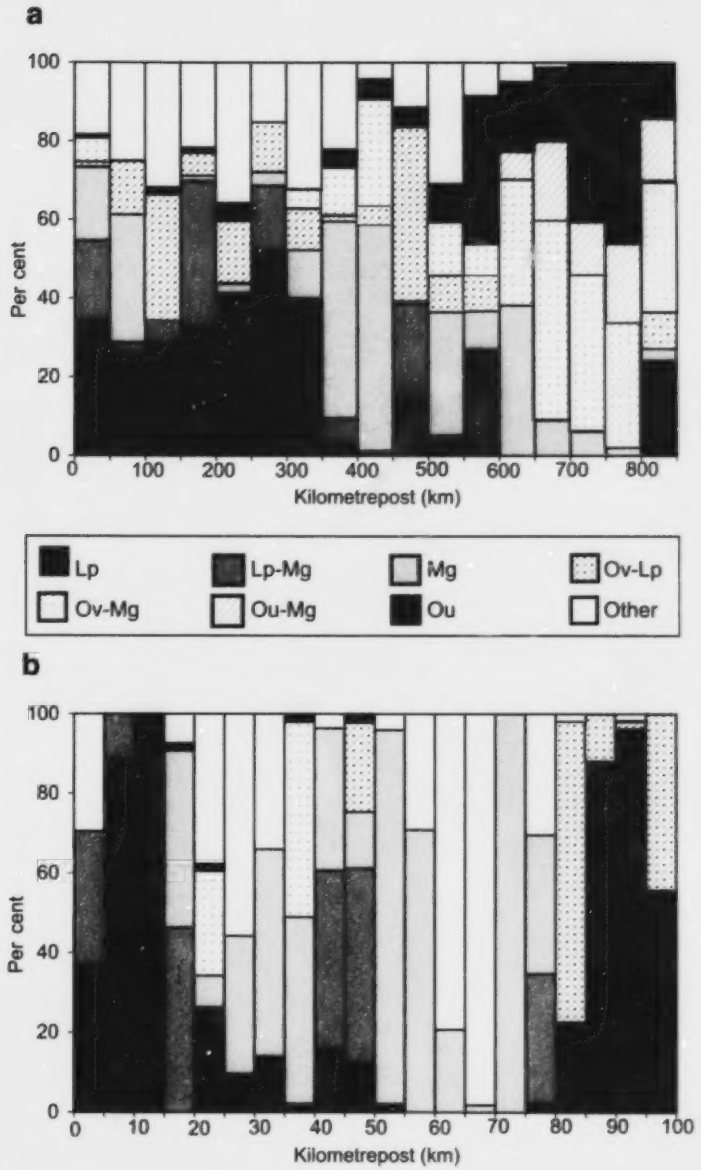


Figure 4. *Distribution of major terrain units.* ***a****) by 50 km intervals along the entire pipeline route, and* ***b****) by 5 km intervals in the first 100 km of the pipeline route. Lp — lacustrine landforms, generally plains, including veneers (alluvial, colluvial, eolian, and glaciofluvial) over lacustrine plain units; Mg — moraine, generally ground moraine, including veneers (alluvial, eolian, glaciofluvial) over moraine units; Lp-Mg — composite/complex of lacustrine and moraine units, also includes veneer of one over the other; Ov-Lp — organic veneer over lacustrine units; Ov-Mg — organic veneer over moraine units; Ou-Mg — composite of organic and moraine units; Ou — organic units: fen, bogs, and fen-bog complexes; Other — alluvial, colluvial, eolian, bedrock, and glaciofluvial units.*

unfrozen with frozen trench bottom; rare case documented where the seasonal freezeback was incomplete but trench bottom was in permafrost.

The ditchwall logs were analyzed to determine the percentage of terrain that could be interpreted as permafrost (conditions F and F/U/F). It should be noted that observations were at times varying from early January to early March. Some underestimation may have occurred since F/U conditions could in fact represent a case where the active layer is deeper than the trench. Active layer thicknesses observed in undisturbed and uncleared terrain of the central and southern Mackenzie valley are, however, generally less than 120 cm (the depth of the trench); *see*, for example, Table 1 and Figure 5 in Nixon (2000). True underestimation more likely could have occurred in areas where the pipeline occupies a previously cleared right-of-way and some deepening of the permafrost table may have already taken place (as discussed below). Although the newly cleared right-of-way had generally experienced one thaw season before trenching, active layer depths were likely still less than 1 m (based on active layer depths of <1 m measured on the right-of-way at the end of the first thaw season after construction at GSC study sites in newly cleared sections of the northern part of the route).

Figure 8a presents the percentage of permafrost terrain by 50 km interval. There is a general southerly decrease in the percentage of permafrost terrain, from a high of more than

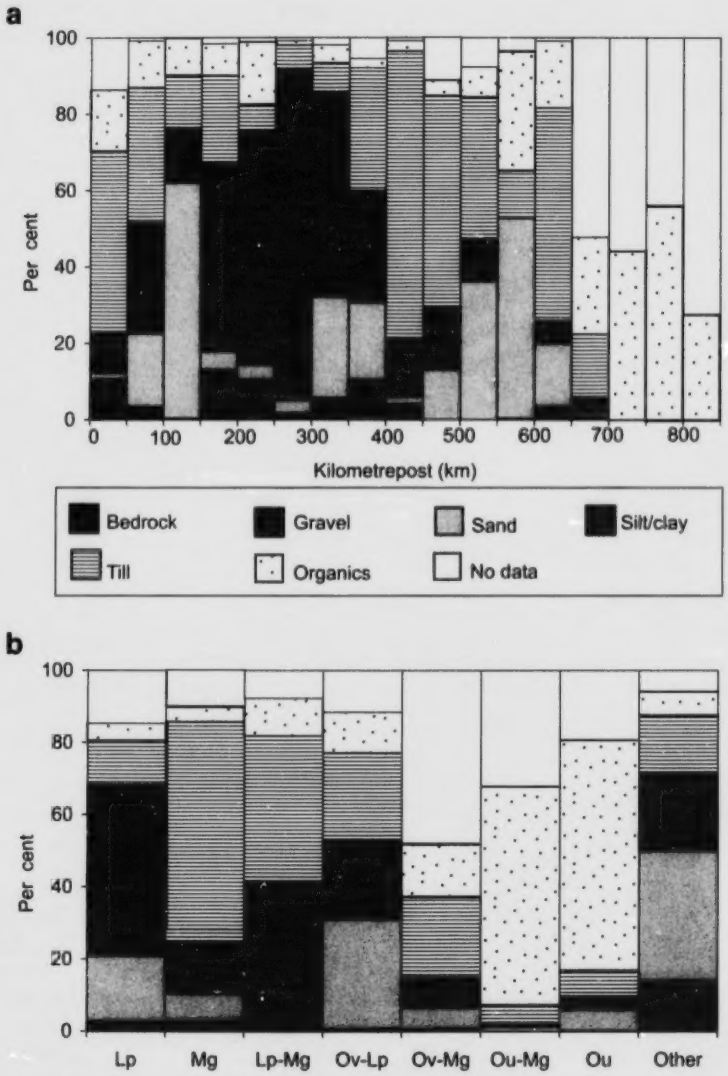


Figure 5. *Distribution of soil groups:* ***a)*** *by 50 km intervals along the pipeline route, and* ***b)*** *within major terrain units for the whole route.*

80% in the first 200 km of the route to a low of less than 30% in the kp 550–650 area. This general trend is consistent with the north-south routing of the pipeline through the permafrost zone (*see* Heginbottom, 2000). South of kp 650, the percentage of frozen ground rises as the elevation increases onto the Alberta plateau (*see* physiographic regions in Fig. 1b). The percentage of permafrost between kp 200 and 300 is much lower than elsewhere in the northern part. This low percentage correlates with a high occurrence of clay and silt (*see* Fig. 5) and is perhaps related to high unfrozen water contents of these materials (which may lead to their being described as unfrozen at temperatures slightly below 0°C). Figure 9 provides a detailed look at frozen ground distribution in the 100–300 km region by 5 km intervals.

Part of the pipeline route selection philosophy was to maximize the use of previously cleared right-of-way, in order to minimize the amount of new permafrost thaw and associated settlement that would be initiated by construction clearing. Thus an examination of permafrost occurrence was also undertaken on the basis of type of right-of-way clearing: 1) 'uncleared' prior to the pipeline clearing and construction, and 2) 'previously cleared': either a) the 20–25 year old CNT

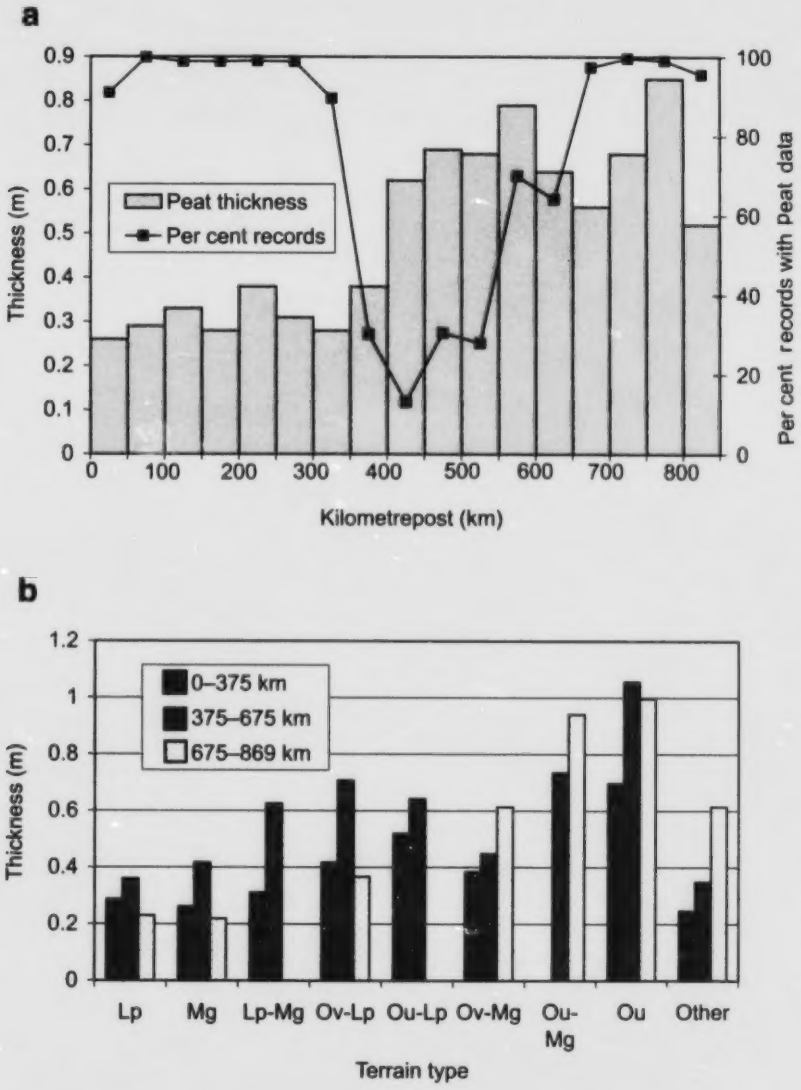


Figure 6. *Average peat thickness:* ***a****) for each 50 km interval along the route; also shown is the percentage of records in each interval with information on peat thickness, and* ***b****) for each major terrain type, dividing the route into three sections: kp 0–375, kp 375–675, and kp 675–869.*

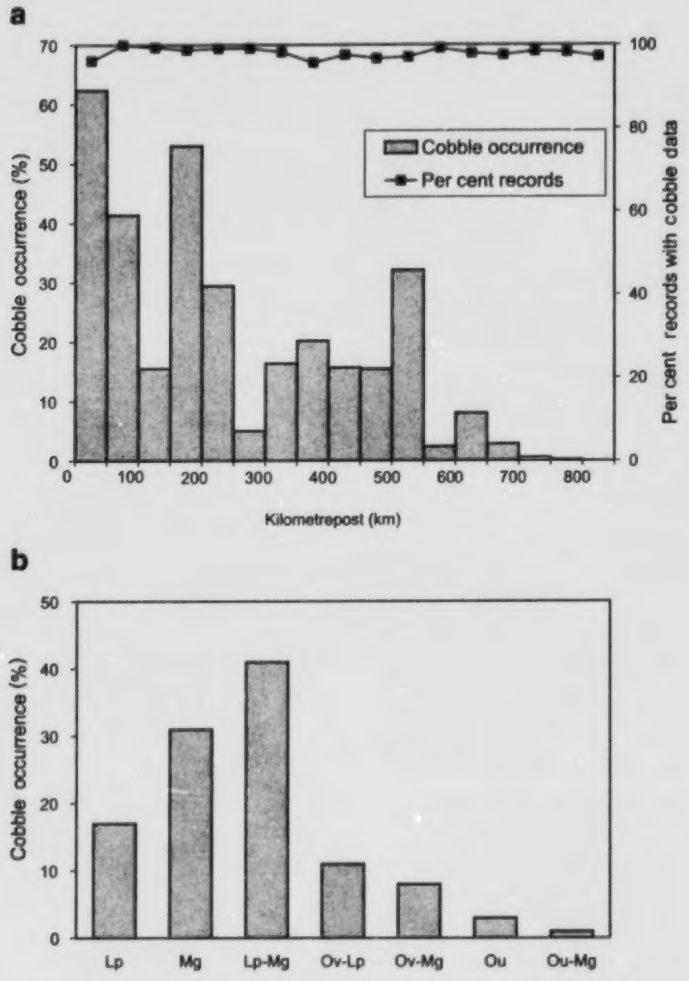


Figure 7. *Summary of cobble occurrence:* ***a****) by 50 km intervals along the pipeline route, and* ***b****) by terrain unit. The percentage of records in each 50 km interval that have information on cobble occurrence is also plotted in a).*

(Canadian National Telecommunications) cutline or b) 'other' (such as seismic lines) clearings widened for pipeline construction. Figure 8b shows the distribution of the various types of right-of-way: approximately 60% of the route was previously cleared in the northern section (kp 0–400) compared to 25% in the south (kp 400–869). Figure 8c shows the percentage of frozen ground in previously uncleared right-of-way only, a distribution similar to that for all right-of-way types in Figure 8a. Permafrost occurrences are only a few percent higher in Figure 8c (compared to 8a) suggesting that the extensive use of previous clearing in the north did not have a major impact on the presence or absence of permafrost in the trench.

The distribution of permafrost within previously cleared right-of-way only is shown in Figure 8d (Note: some intervals are blank because they only have newly cleared right-of-way). The general pattern of reduced percentage of frozen ground compared to Figure 8c probably reflects the impact of previous clearing on the degradation of permafrost. The greatest impact has occurred along the southern part of the route (but only a small percentage of the route in the south used previous clearings, and a large percentage of these clearings would have been unfrozen in any case).

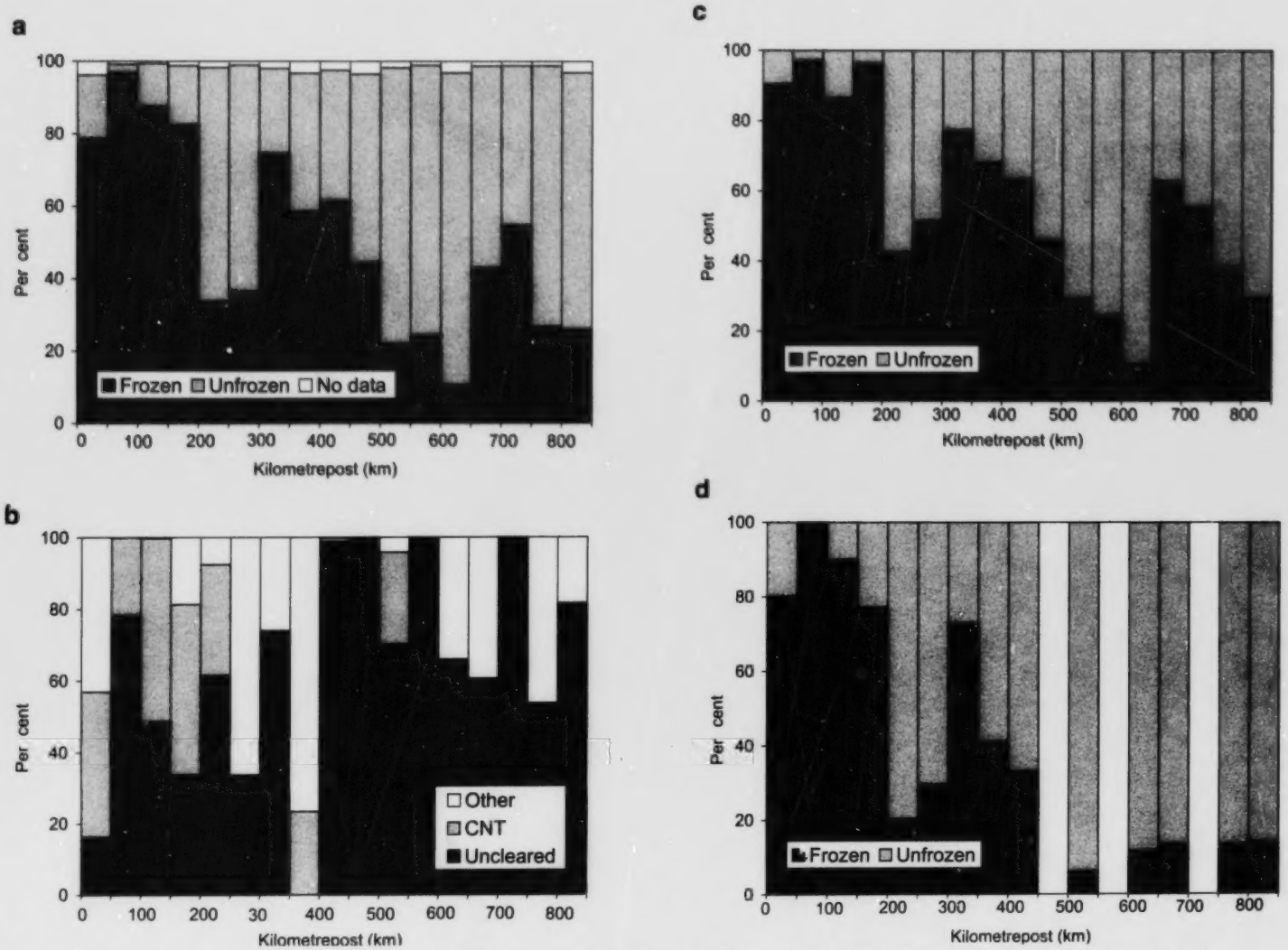


Figure 8. *a) Frozen ground distribution by 50 km intervals for all types of right-of-way; b) Distribution of right-of-way types in each 50 km interval along the entire pipeline route. Uncleared = right-of-way that had not been previously cleared, i.e. newly cleared for pipeline construction; CNT = the right-of-way followed the Canadian National Telecommunications clearing that had been made in the 1960s and was originally 12 m wide; OTHER = right-of-way had been cleared, generally to a width of 6 m or less for other purposes such as seismic lines; c) Frozen ground distribution in previously uncleared right-of-way, by 50 km intervals. d) Frozen ground distribution in all types of previously cleared right-of-way by 50 km intervals. Note: some intervals are blank since they have only newly cleared right-of-way.*

Visible ice distribution

Observations on the type and amount of visible ice in the trench were recorded for the first 335 km of the route. Values entered in the database were those for the trench bottom whenever possible, generally rounded off to the nearest 5%. Figure 10 plots these values as a function of kilometrepost; the data have been separated into observations for 'new right-of-way' and 'old (previously cleared) right-of-way'. In Figure 10, there is considerable scatter in the data, reflecting the different terrain units and soil textures. The scatter is similar for both types of right-of-way. The average of all values is around 20%. A trend line through all the data reveals a slight decrease in percentage of visible ice from north to south.

An examination of the kp 0–335 data, as a function of terrain type and section of the right-of-way was also undertaken. This analysis was based only on those records where the trench bottom was described as frozen (F) and where a description was given for the ice type. These represent a total of 1937 records. Eighty-five per cent of these records contain

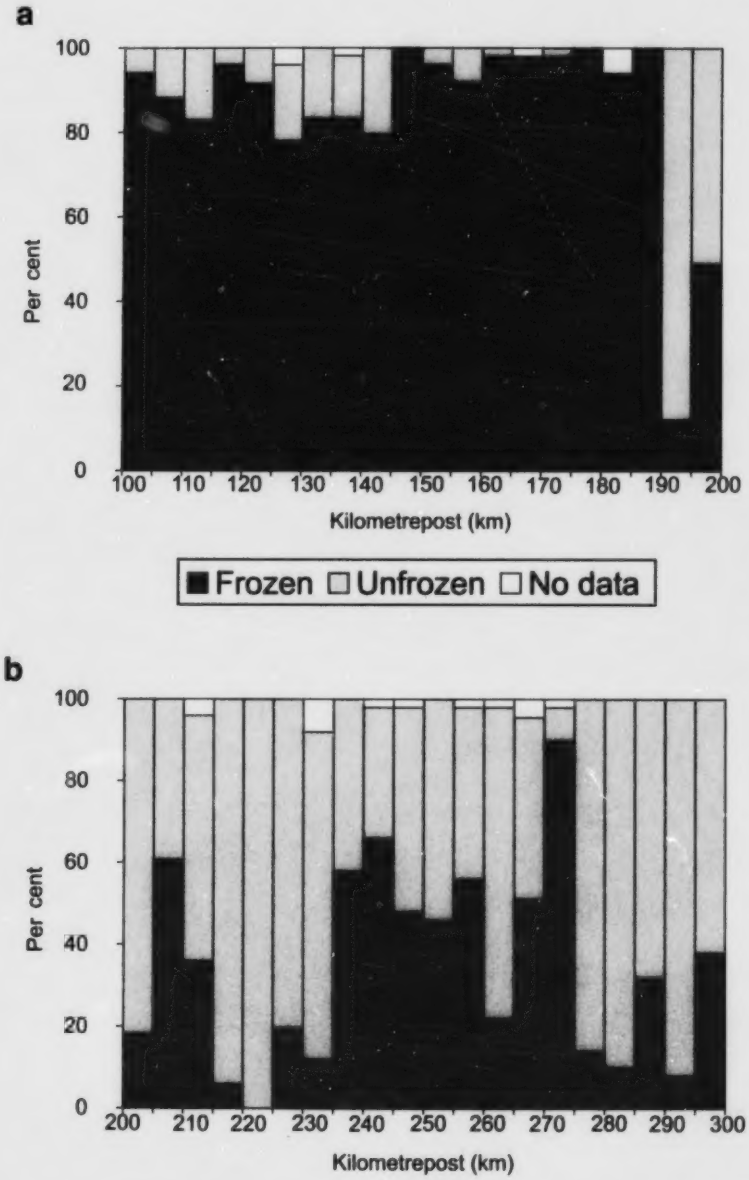


Figure 9. *Frozen ground distribution by 5 km intervals,* ***a)*** *kp 100–200* ***b)*** *kp 200–300.*

a percentage value for ice. Thirteen per cent of the records have no value logged but are described as having no visible or excess ice; a value of 0% was therefore assumed for these records. A value of 0% was also assumed for the remaining 2% of the records, although these could represent instances where a per cent ice value was not recorded. The average visible ice content (in per cent) was then calculated for each of the major terrain units, in each of three areas (kp 0–100, 100–200, and 200–335). The results are shown in Figure 11, distinguishing between previously cleared and newly cleared right-of-way. (Note: terrain units which were grouped elsewhere under 'other' have been identified here and are as follows: Go – glaciofluvial, Ed – eolian dune, Cm – colluvial, Af – alluvial fan, Ov-Go – organic veneer over glaciofluvial, Ou-Go – mixture of organic and glaciofluvial, Ou-Lp – composite of organic and lacustrine.)

The ground-ice distribution as a function of terrain type and section of right-of-way reveals that the weak general trend shown in Figure 10 is no longer apparent when examining a specific terrain unit, for example Lp or Lp-Mg. Ice contents are generally lower in the 100–200 kp range than in either the 0–100 kp range or the 200–335 kp range. The average ice content in new clearings is more likely to be higher than that of previously cleared right-of-way in the same terrain unit, but it is often lower or the same. Clearing of the pipeline right-of-way took place one year prior to construction, and the resulting increase in active-layer depth could have decreased the amount of ground ice in the trench, especially for newly cleared portions. Active layer depths measured on the right-of-way at the end of the first thaw season after construction at newly cleared GSC study sites in the northern part of the route were however generally less than 1 m. Thus ground-ice conditions in the base of the trench were likely 'original' for segments of newly cleared right-of-way.

Frozen-unfrozen transitions

The possibility of differential heave or settlement across interfaces between frozen and unfrozen ground is a principal concern for buried pipelines in permafrost terrain. Knowledge of the number of frozen-unfrozen transitions was thus an important consideration in the design of Norman Wells pipeline. Continuous electromagnetic surveys were conducted along the right-of-way prior to construction (Kay et al., 1983)

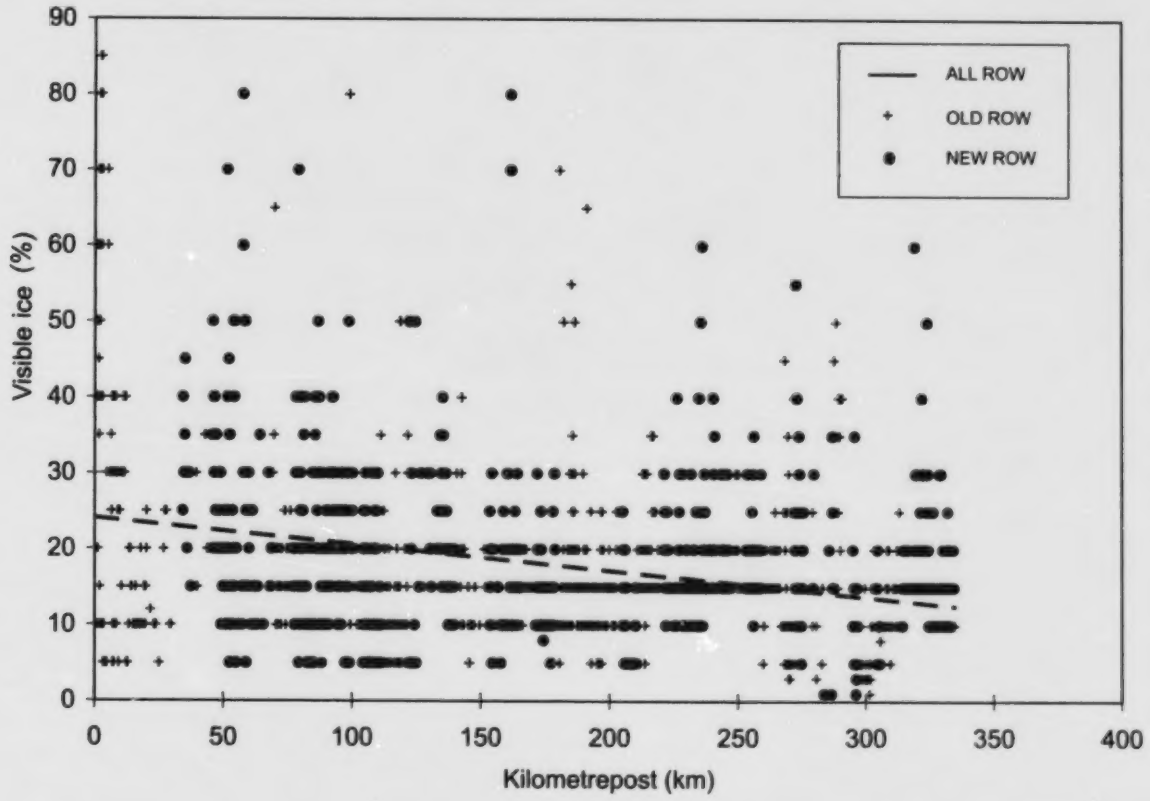


Figure 10. *Visible ice content recorded in the trench bottom for the first 335 km of the pipeline route. Observations in newly cleared right-of-way are distinguished from those in previoulsy cleared right-of-way. Trend line is for all data points.*

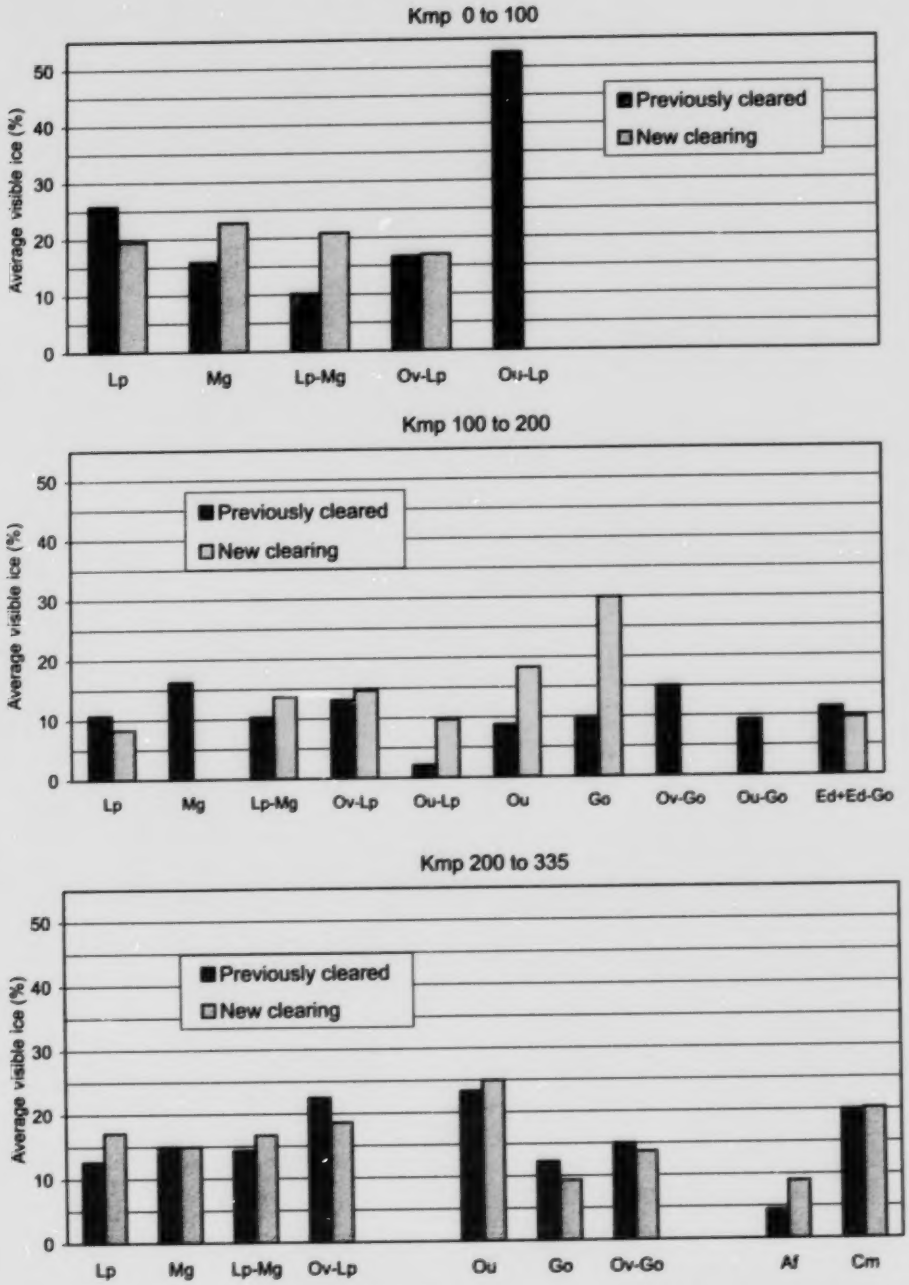


Figure 11. *Average visible ice content for each major terrain unit, shown separately for each of previously cleared and newly cleared right-of-way types. Calculations of the average are based only on those records in the database that have an entry in the ice description field as well as the ice content field.*

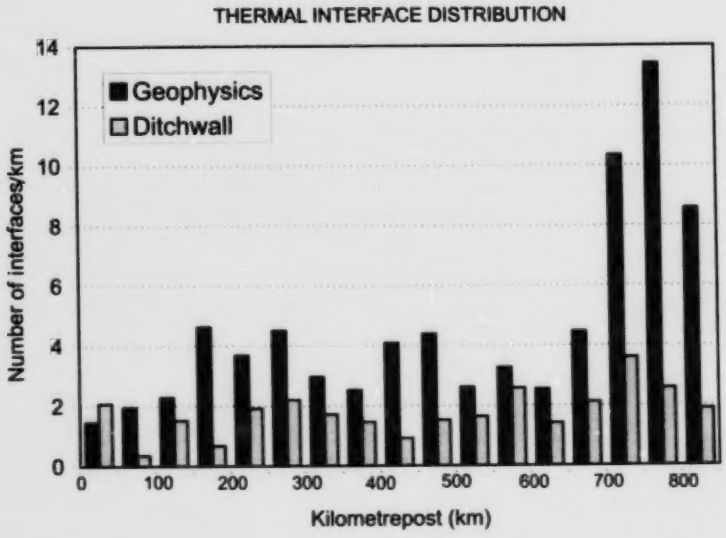


Figure 12.

Average number of frozen/unfrozen interfaces per 50 km interval, as recorded in the ditchwall database and the geophysical surveys (EM).

to determine the presence or absence of frozen conditions in the top 10 m of the ground. The number of interfaces per 50 km interval recorded by the geophysics is compared to the number recorded in the ditchwall database in Figure 12. There is reasonable agreement between the two data sources in the first 700 km; an average of two interfaces per kilometre in the ditchwall data compared to an average of three in the geophysics. South of kp 700, however, the geophysical data show an average of ten interfaces per kilometre while the ditchlogs show only three. Nixon et al. (1991) examined the frozen-unfrozen transitions in the ditchwall database, their frequency along the route, relationship to terrain type, and the length of frozen and unfrozen segments. They also discussed the possible significance of the differences between the ditchwall and geophysical data. The discussion below adds to this analysis and is based on a more complete version of the digital ditchwall database.

Analysis of the ditchlog data, by terrain type, revealed that the average number of interfaces per kilometre is highest in the organic terrain (typically 4–5/km). Lacustrine plains averaged 3/km, while till plains average 1/km. In the northern end, frozen sections are longer, whereas the reverse is true in the south. The lengths of frozen and unfrozen sections were examined in the south, where the number of interfaces was the greatest, using data from kp 500–870 and 5 km intervals. The average length of an unfrozen section is 1 km in the ditchlog data and 0.75 km in the geophysical data. The average length of a frozen section is 0.3 km in the ditchlogs and 0.1 km in the geophysics.

The discrepancies between the geophysical and ditchwall data may be related to several factors:

1) The shallow depth of observations of the ditchlog compared to the geophysics.

2) The geophysical surveys were conducted to obtain information beneath the depth of pipe burial, since this was of greatest importance for pipe design. The condition 'frozen' was assigned in the geophysical surveys when at least 3 m of the top 10 m of ground was interpreted as being frozen. If permafrost was thin (<3 m) this would be interpreted as unfrozen, thus giving rise to more interfaces in the geophysical database than in the ditchwall. For example on the Alberta plateau, the total length of unfrozen sections interpreted by geophysics is greater than in the ditchwall logs, which would agree well with the thin permafrost conditions being interpreted as unfrozen. From a pipeline design perspective, these geophysical interfaces in the south between unfrozen or thin (<3 m) permafrost and thick (>3 m) permafrost, have important implications for differential pipe movements.

3) Some of the differences in geophysical response may be related to textural, moisture, or stratigraphic transitions rather than to frozen/unfrozen interfaces; however, these factors were considered in the data interpretation.

The differences were further examined by interpretation of airphotos taken shortly after construction in 1985. The specific sections examined were those south of the Mackenzie River crossing (kp 528–869). This approach focused on identifying fens and other readily recognizable unfrozen areas crossing the right-of-way (i.e. unfrozen and unrelated to construction). The aim was to obtain a minimum for the number of frozen/unfrozen transitions, since no attempt was made to identify transitions in mineral soil or between bogs and mineral soil. The average number of interfaces per kilometre identified from the airphotos was two. Between kp 700–869, where the greatest differences were noted between the ditchwall logs and the geophysics, the average was also 2 interfaces/km. Since the airphoto results represent a minimum number of interfaces, the actual number of interfaces

between unfrozen and permafrost terrain (regardless of thickness of permafrost) is likely somewhere between the ditchwall and geophysical values.

SUMMARY

An extensive record of permafrost and terrain conditions in the central and upper Mackenzie valley has been digitized in the ditchwall database from the Norman Wells pipeline construction. This database is unique in providing a detailed continuous transect through the discontinuous permafrost zone of soil types, peat thickness, ground thermal conditions, and visible ice content in the top 1.2 m of the ground. It allows the quantification and analysis of trends along the north-south corridor, as a function of distance, terrain unit, and type of clearing. The database provides base-line geoscience information for the assessment of the impact of climate change on the physical environment. Such detailed data over a wide area are rare, yet important to the validation of regional-scale permafrost prediction models such as that presented later in this volume (*see* Wright et al., 2000).

ACKNOWLEDGMENTS

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Shoreline permafrost along the Mackenzie River

L.D. Dyke[1]

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Abstract: Shorelines mark one of the most abrupt changes in ground temperature in permafrost regions. The ground temperature in the vicinity of shorelines is controlled by the contrast in mean annual temperature between the particular water body and the adjacent land. Erosion or sedimentation along shorelines complicates ground temperatures by moving the boundary between relatively cold land and warmer water. A river channel migrating due to erosion on the outside of a bend introduces the relatively warm water condition to frozen ground, causing thaw to begin. At the same time, sediment is deposited on the opposite shore as the migration progresses. This results in ground temperatures again being determined directly by air temperature. Depending on water temperature and shoreline vegetation, temperature contrasts can range from about 4°C to over 10°C.

Résumé : C'est dans les berges des régions pergélisolées que l'on observe les changements de température les plus marqués dans les sols. La température du sol près des berges dépend de l'écart qui existe entre la température annuelle moyenne d'une étendue d'eau donnée et celle des terres adjacentes. L'érosion ou la sédimentation le long des berges modifie la répartition des températures dans le sol en déplaçant la limite entre les sols relativement froids et l'eau plus chaude. La migration d'un chenal fluvial causée par l'érosion de la berge extérieure d'un méandre met l'eau relativement chaude du cours d'eau en contact avec le gélisol, causant le dégel de celui-ci. Pendant que se déroule ce processus, des sédiments se déposent sur la berge opposée. Une fois de plus, les températures du gélisol dépendent directement de la température de l'air. Compte tenu de la température de l'eau et de la végétation de la berge, l'écart de température peut varier d'environ 4 °C à plus de 10 °C.

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INTRODUCTION

In permafrost regions, shorelines produce the most abrupt contrasts in ground temperature. Furthermore, shorelines change position because of erosion, sedimentation or sea-level change. This paper describes the probable effects of shoreline movement on permafrost distribution at sites in the Mackenzie valley. The locations of sites discussed are shown in Figure 1.

THERMAL REGIME ALONG THE SHORELINE

Where water depth steadily decreases toward shore, permafrost thickens landward from the point where winter ice comes in contact with sea, lake, or river bed sediments sufficiently long to freeze them perennially. As the water depth continues to decrease, the duration of ice contact increases, producing lower ground temperatures and thicker permafrost. Where the bottom profile is steep, these changes are abrupt. Figure 2 shows the distribution of average annual temperatures across this zone of ice contact with bottom sediments. This example represents the Mackenzie Delta coast, where the greatest range of ground surface temperature in the Mackenzie valley is likely to be encountered. Farther south the minimum temperature is warmer and the thickness of winter ice is less.

Figure 1. *Lower Mackenzie River showing locations of sites discussed.*

Horizontal distance is not shown in Figure 2 because the width of the ice contact zone may vary from metres to kilometres, depending on the gradient of the bottom adjacent to the shore. Where the width of the ice contact zone is considerably greater than the ice thickness, heat flow from water to land is insignificant and ground temperatures are governed only by the average annual temperature beneath a given ice thickness; however, if the nearshore slope is steep (river scouring can produce channel sides steeper than 45°), then the width of the zone is reduced sufficiently so that a significant lateral thermal gradient exists. Contours of average annual temperature (isotherms) radiate from the shoreline (Fig. 3) and the pattern of these isotherms will be determined by the thermal equilibrium established between land and water temperatures. The extremes of shoreline temperature distribution that are likely to be found in the Mackenzie valley where permafrost distribution is in equilibrium with water and land temperatures is shown in Figure 3. A land-to-water contrast of -8°C to 2°C would be representative of the outer Mackenzie Delta and Beaufort Sea coast (Fig. 3a) and a contrast of -2°C to 4°C would be generally typical of any inland, or well vegetated river bank (Fig. 3b). With increasing depth, the lateral thermal gradient ceases to be greater than the geothermal gradient and the isotherms become horizontal.

MIGRATING CHANNELS

The most dynamic shoreline changes in the Mackenzie valley result from river-bank erosion and sedimentation along the Mackenzie River, its distributaries in the Mackenzie Delta, and tributaries. Scouring on the outside of river bends and sedimentation on the inside results in lateral migration of the channel. Along most of its length, the Mackenzie River is

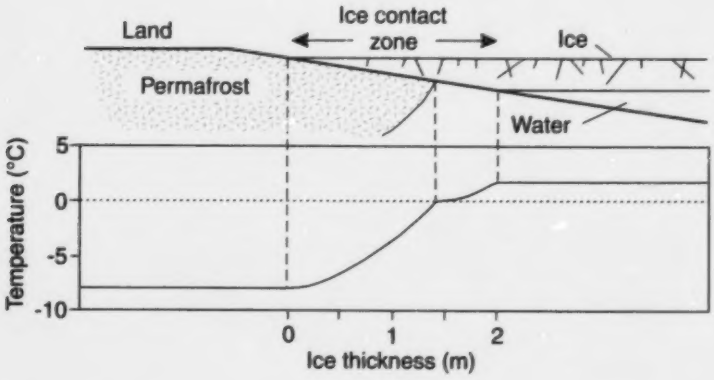

Figure 2.

Distribution of average annual temperatures across the nearshore zone in contact with winter ice, Mackenzie Delta front. Snowdrifts will locally increase temperature.

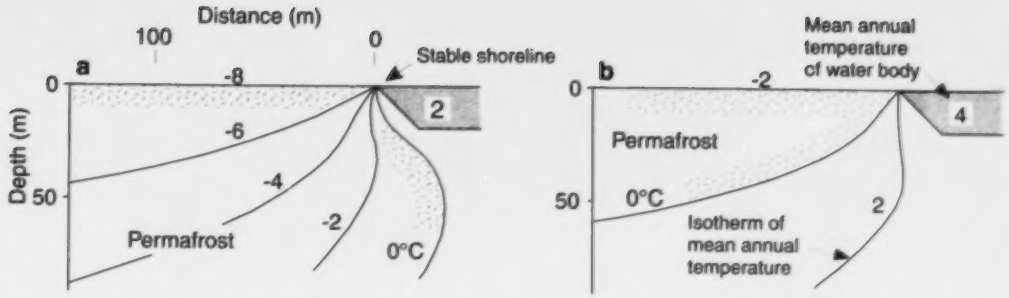

Figure 3. *Average annual ground temperatures adjacent to a laterally stable shore for:* ***a)*** *a temperature contrast between land and water of -8°C to 2°C typical of the Mackenzie Delta front, and* ***b)*** *a contrast of -2°C to 4°C typical of much of the Mackenzie River valley.*

confined within a generally narrow valley bottom. Only in the 100 km section immediately upstream from Tulita (formerly Fort Norman) and in the Mackenzie Delta, is the valley bottom sufficiently wide to permit development of high-amplitude meanders. Bends that are entirely within floodplain sediments migrate most quickly. Migration rates are highly variable but average a few metres per year in the Mackenzie Delta (Lapointe, 1986) and can reach 10 m/year where scouring is strongest. Tributary channel migrations can be faster (*see* Brooks, 2000).

The effect of channel migration is to transport the river heat source across already frozen sediments and thus upset the thermal equilibrium between land and water. Smith and Hwang (1973) modelled this process for a channel in the Mackenzie Delta and show good agreement between measured and predicted ground temperatures. Scouring on the outside of bends continually removes thawed sediment, thereby keeping permafrost at or close to the bed of the channel. With continued channel migration, downcutting will extend to the depth of the channel (thalweg) causing permafrost to extend beneath the river. Eventually the rate of melting of permafrost beneath the river bottom will exceed the removal of sediment, resulting in a thawed zone beneath the channel bottom. Along the opposite side of the thalweg, sedimentation begins along the river bottom and the channel shallows; however, thawing slows because of the increasing distance between the river bottom and the thaw front. Although this process is likely ubiquitous along migrating channels in permafrost regions, few specific cases are documented (e.g. Hollingshead et al., 1978). Migration ultimately results in emergence of the river bottom to form a subaerial bar and re-establishment of permafrost. The size of the resulting thawed zone (talik) beneath the river will depend primarily on migration rate of the channel, the temperature regime on the accreting side of the channel, and the size of the channel.

THE ROLE OF VEGETATION

After climate, vegetation plays the next most important role in controlling the re-establishment of permafrost on the accreting side of channels. The insulating effect of snow trapped by vegetation cover greatly ameliorates winter cold at the ground surface. The expected average annual ground surface temperatures corresponding to different types of vegetation across aggrading shores for three Mackenzie valley settings are shown in Figure 4. Shown in Figure 4a are the temperatures as cold as -8°C at the front of the Mackenzie Delta which reflect the influence of the Beaufort Sea and unvegetated bars. On land only sedge meadows develop. This limits snow entrapment and produces an average annual ground surface temperature of about -5°C. Inland, wide channels are still associated with unvegetated bars but sedges are succeeded by willows (Fig. 4b). Willows trap more snow, raising the ground temperature to about -2°C (Fig. 4b). Banks on the lee side of channels and narrower secondary channels, also offer protection for snow accumulation. Unvegetated bars in these locations may become the warmest subaerial environment in any area because of their burial by the resulting snow drifts (Fig. 4c). Smith (1976), in a detailed study of subsurface temperatures beneath an aggrading point bar in the central Mackenzie Delta, found the temperature of the bar surface to average about -1°C. His study site shows a succession of sedge to willow, ending in spruce forest with a lowering of temperature to -4°C because of summer shading and the interception of some snowfall by the forest canopy in winter. The data for preparing Figure 4 is derived mainly from the Mackenzie Delta and therefore temperatures will be expected to increase gradually with increasing distance south; however, temperatures for vegetated alluvial sites in the Norman Wells area (A. Judge, pers. com., 1995) are only about 1°C warmer than equivalent sites in the Mackenzie Delta, suggesting that Figure 4c is broadly applicable to a considerable part of the Mackenzie valley.

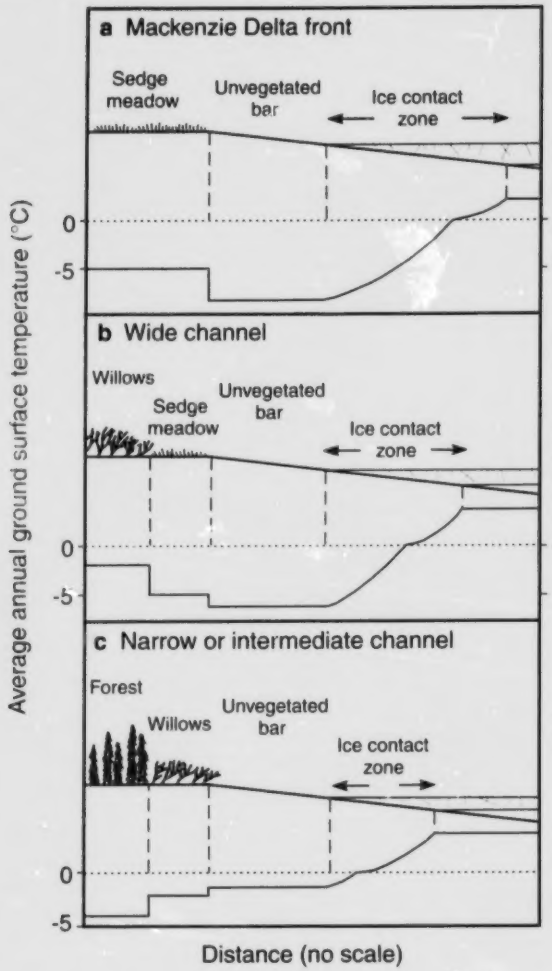

Figure 4. *Schematic drawing of the average annual ground surface temperatures characterizing three riverbank environments within the Mackenzie Delta:* ***a****) Beaufort Sea coast at the Mackenzie Delta front,* ***b****) wide channels and major distributaries,* ***c****) narrow and intermediate channels.*

PREDICTION OF THAWING BENEATH MIGRATING CHANNELS

All permafrost distributions presented in this chapter are predicted by using the one-dimensional heat-flow equation modified to account for complete melting of ground ice at 0°C (Neuman solution; Jumikis (1977)). These predictions agree with limited observations recorded in the course of pipeline crossing explorations where permafrost boundaries are observed to be close to channel sides that are undergoing scouring (petroleum company observations included in Traynor and Dallimore (1992)); however, the observations are too widely spaced to locate boundaries precisely so that theoretical predictions have been used to determine sensitivity to the conditions noted above. A single set of sediment conditions and thermal properties is used in all examples: unfrozen thermal conductivity = 1.8 Wm/°C, frozen thermal conductivity = 3.0 Wm/°C, unfrozen volumetric heat capacity= 2.0 J/cm^3°C, frozen heat capacity = 1.2 J/cm^3°C, and water content by weight= 20%, giving a latent heat content of 117 J/cm^3.

A generalized channel profile with a width typical of major Mackenzie Delta distributaries (e.g. *see* Fig. 7) but similar in shape to cross-sections elsewhere along the river is shown in Figure 5. Thawing estimates are based on an average annual water temperature for the Mackenzie River of 4°C. Water temperatures rise slightly with distance downstream from the source at Great Slave Lake but do not vary more than 1°C from this value (MacKay and Mackay, 1974). The depth of thaw is directly dependent on this temperature and the length of time over which it prevails. After sediment accretion raises the river bottom into the ice contact zone, thawing stops and refreezing from the surface occurs in response to a progression of thermal conditions as represented in Figure 4. The predicted talik for migration rates of 1 m/year and 3 m/year with a vegetation succession on the aggrading shore from sedges to willow is shown in Figure 5. These rates are rarely exceeded except in the lower part of the Mackenzie Delta. In both cases the talik extends through permafrost. In all examples, the vegetation succession is assumed to follow the migration of the river concomitantly, thereby controlling the average annual ground temperature.

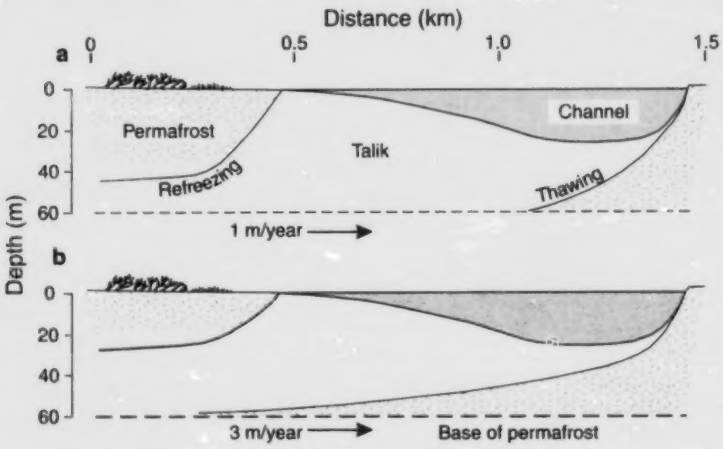

Figure 5.

Predicted talik developed beneath a major distributary channel having migration rates of ***a****) 1 m/year and* ***b****) 3 m/year. In both profiles, refreezing decelerates as willows with associated warmer ground temperatures follow the channel migration. Thawing decelerates as the frost table falls below the channel thalweg. Increasing migration rate increases the proportion of channel underlain by permafrost. Average annual ground temperatures used to calculate refreezing are shown in Figure 4b.*

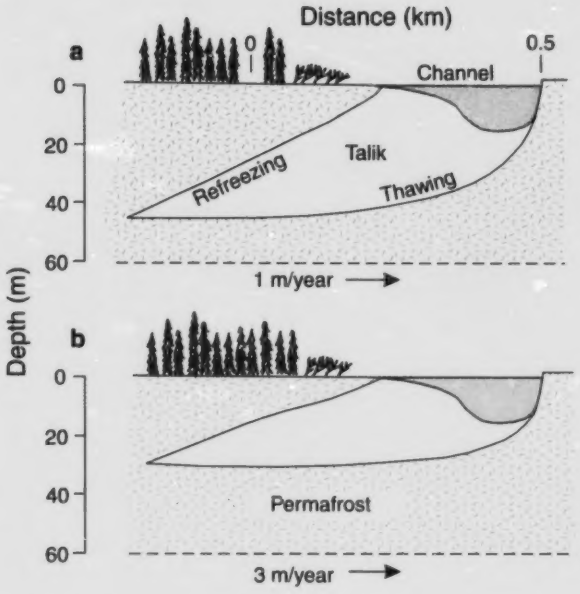

Figure 6. *Predicted talik developed beneath a secondary channel having migration rates of 1 m/year* ***a****), and 3 m/year* ***b****). Because the channel is much narrower than in Figure 5, thawing decelerates more quickly and refreezing begins sooner, resulting in a talik which does not breach permafrost. Ground temperatures for refreezing are also lower because of the establishment of forest. Average annual ground temperatures used to calculate refreezing are shown in Figure 4c.*

For a laterally stable channel in the Mackenzie Delta, Smith (1976) indicated that a through-permafrost talik will exist for any channel wider than about 70 m; however, migration increases the width of channel required to produce a through-talik. A 200 m wide channel with a willow-to-spruce succession approximating conditions at the site studied by Smith (1976) is shown in Figure 6. In this case, the talik refreezes well before it reaches the base of permafrost for both the 1 m/year and 3 m/year migration rates. Even for the 1 km wide channel shown in Figure 5, with 3 m/year of migration, thaw does not penetrate permafrost beneath the channel. On the other hand, repeated passages of migrating channels over the same location will limit the maximum depth of the base of permafrost.

PERMAFROST THICKNESS IN THE VICINITY OF MIGRATING CHANNELS

Migration of the bend in a major channel shown in Figure 7 has been determined by tracing previous channel positions visible on this aerial photo. Channel positions prior to 1950 can be traced as faint scars in the medium grey-toned, willow-covered area landward of the present point bar. Their age can be roughly estimated by assuming that the migration rate for the 35 year time span documented in Figure 7 (about 6 m/year) is representative of the past. The boundary of dark forest adjoining the willow area outlines a former position for the inside of the bend. Although the traces of former channel positions lack the same channel curvature as the present bend, the old channel scars mark incremental positions that appear to be directly related to movement of the bend from this prior position. Therefore it is concluded that the present willow-covered area has aggraded at about the same rate as the outer bend.

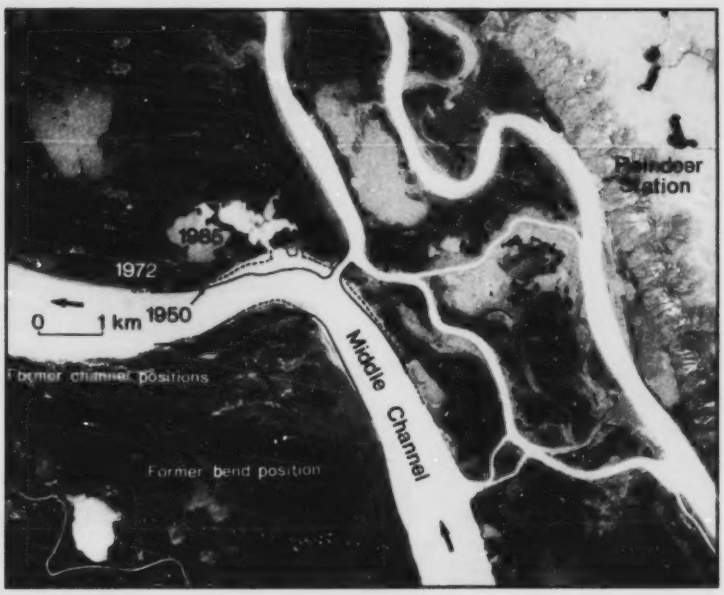

Figure 7.

A 1985 aerial photo showing migration of a major bend in Middle Channel, east central Mackenzie Delta. Former shorelines for 1972 (dashed) and 1950 (solid) are indicated. NAPL A26726-19

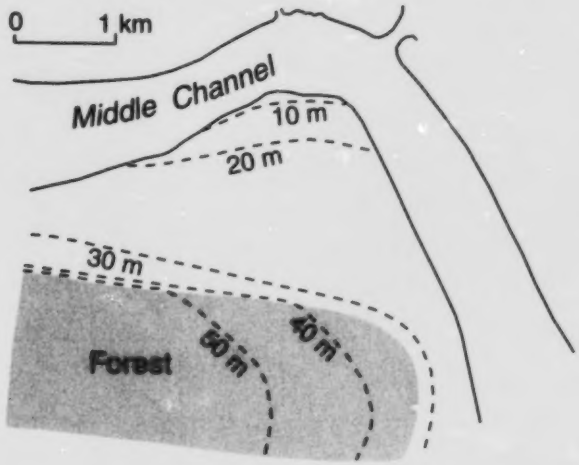

Figure 8. *Predicted permafrost thickness (in metres as shown by dashed lines) beneath the zone of point bar deposits for the major bend in Figure 7. The thickness contours parallel former channel margins. The warming effect of lakes is not taken into account.*

With an estimate of the age gradient across the willow-covered area, thickness of permafrost can be calculated. Permafrost growth will be controlled by the same succession of bank conditions as shown in Figure 5 with willow determining the average annual ground temperature until the forested area is reached; however, the time of forest establishment is not known, thus making the period that the forest determined the ground temperature difficult to include. For the sake of illustration, it is assumed that the forest was established when the channel position coincided with the bend outlined by the forest. Earlier channel positions within the forested area appear to be truncated at the forest edge, suggesting that an abrupt change in permafrost thickness across this boundary will occur. Notwithstanding these difficulties, Figure 8 has been prepared so that permafrost thicknesses over the entire area can be shown. It is intended only as an indication of the variation of permafrost thickness possible in an alluvial environment.

ISLANDS

Throughout much of its length, the Mackenzie River ranges from 1 km to 5 km in width. Islands and unvegetated bars commonly exist in the wider reaches. These islands and bars migrate downstream, as erosion removes the upstream end and sedimentation adds to the downstream end. Upstream

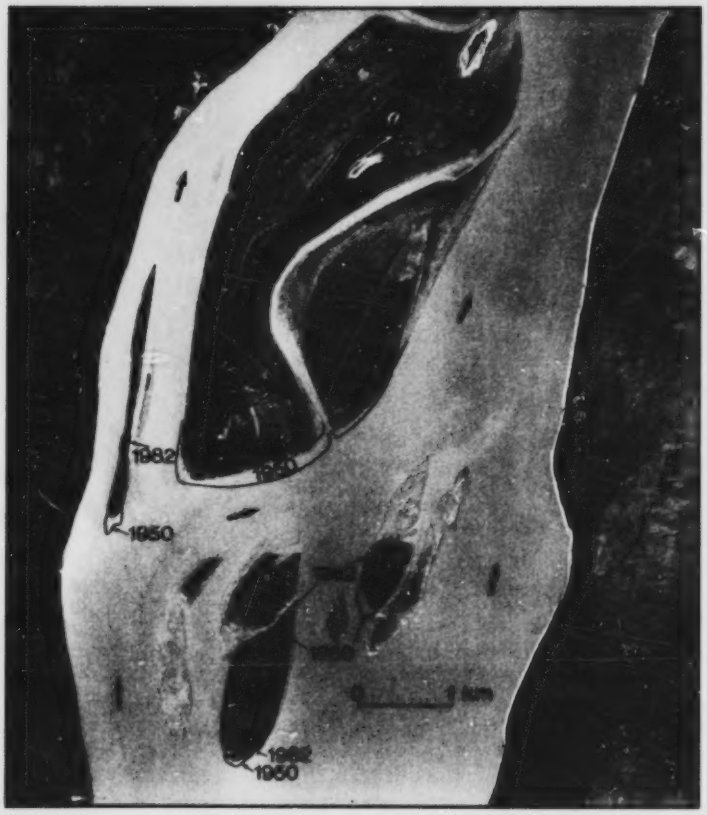

Figure 9.

A 1967 aerial photo of a major alluvial island and smaller island bar complex about 90 km downstream of Fort Good Hope. Channel currents are indicated by arrows. The upstream end of the major island is undergoing erosion by the current moving from left to right against the shore. Shoreline positions for vegetated islands in 1982 and 1950 are shown. NAPL A19947-7

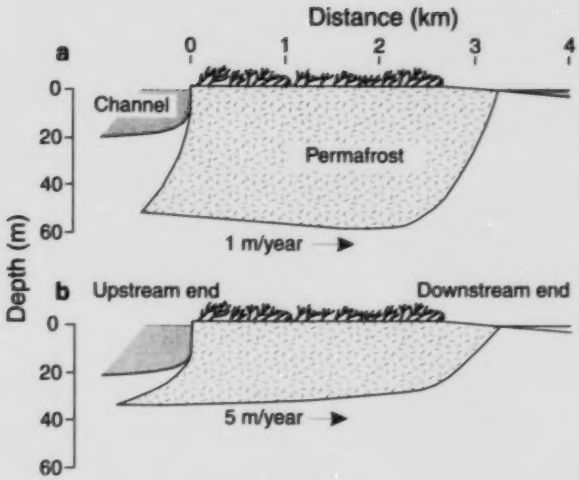

Figure 10. *Cross-sections showing predicted permafrost distribution beneath migrating islands in the Mackenzie River for channel migration rates of 1 m/year* ***a****) and 5 m/year* ***b****). The temperature distribution used ranges from 4°C for the river to -4°C on the downstream bar, warming to -1°C for the willow-covered majority of the island. Increasing migration rate reduces the maximum thickness of permafrost which can develop which in turn reduces the amount by which permafrost can underlie the migrating channel.*

migration, where sedimentation adds to the upstream end and channel erosion removes the downstream end, is also possible. The direction in which erosion takes place depends on the positioning of a channel and shoal water with respect to an island. Downstream erosion is occurring in the location shown in Figure 9 where the main channel flows from the bottom left side of the figure to the top right side, subjecting the upstream end of the large island to significant erosion.

The theoretical distribution of permafrost beneath the longitudinal cross-section of a large island such as that shown in Figure 9 is illustrated in Figure 10. Permafrost aggrades rapidly beneath the downstream unvegetated end of the island, which is migrating downstream, but then diminishes in growth rate in the upstream direction with establishment of vegetation. On a large island, rate of migration fails to alter greatly the amount that permafrost extends beyond the shore of the island beneath the river. Deeper growth permitted by slow migration is compensated by a greater depth of thaw occurring close to the upstream or eroding end.

AN EXAMPLE OF TEMPERATURE CHANGE

As a final example of change in the alluvial environment, the effect of erosion and sedimentation on ground temperatures beneath unvegetated bars is examined. Lack of subsurface data for unvegetated bars in the Mackenzie valley restricts discussion to the Mackenzie Delta front where considerable data are available. This area also provides one of the most dynamic sedimentary environments of the entire river system. Unvegetated bars are large and are susceptible to open ocean wave action and delta-channel migration. This, combined with the coldest average annual air temperatures in the Mackenzie valley, gives the delta front the widest range of ground temperatures and the most rapidly changing permafrost distribution in the Mackenzie valley.

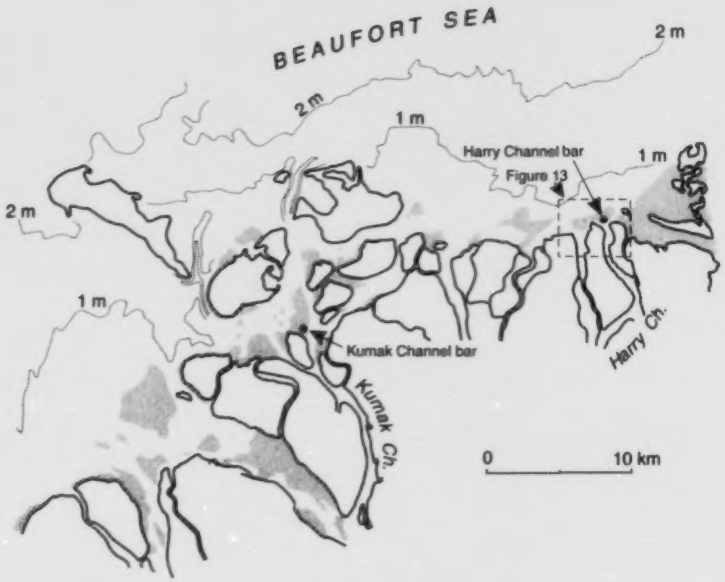

Figure 11.

Mackenzie Delta front showing areas exposed as unvegetated subaerial bars between 1950 and 1985 (grey shading). Also shown are locations of temperature cables giving temperature profiles shown in Figure 12. Dotted lines are bathymetric contours. Location of Figure 13 is shown as dashed rectangle.

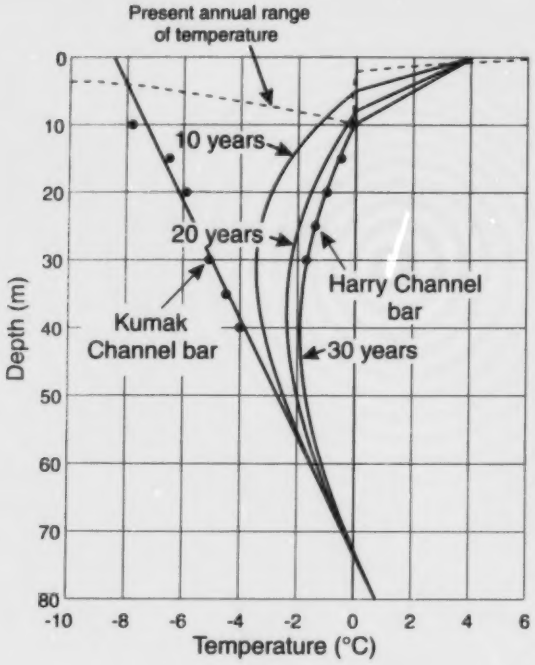

Figure 12.

Temperature profiles beneath two bars at the Mackenzie Delta front assumed to have formed at the same time. Inundation of Harry Channel bar is simulated by suddenly warming the surface temperature to 4°C. The response of the ground temperature 10, 20, and 30 years after this change is shown as a warming from the initial temperature profile represented by the profile beneath the Kumak Channel bar. Dashed lines indicate the 1992 envelope of temperature variation for the Harry Channel bar.

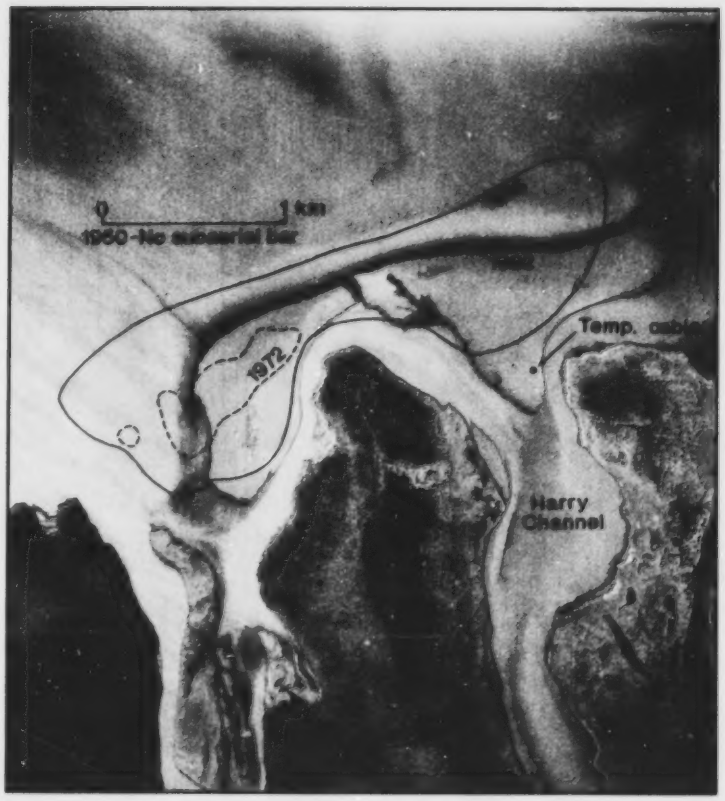

Figure 13.

Development of a bar at the mouth of Harry Channel (east side of the Mackenzie Delta) with former outlines for 1985 and 1972 shown on a 1992 aerial photo. No bar is present on aerial photography for 1950. Location of the Harry Channel bar temperature cable used in Figure 12 is indicated.

The dynamic nature of the delta front is indicated by the variability of sub-bottom temperatures in the apron of bars and shallow water (up to 1 m depth) that extends several kilometres offshore (Fig. 11). Subaerial bars are subject to cold ground temperatures because of windswept exposure and a lack of snow-trapping vegetation. Measured temperature profiles located beneath two bars assumed to have formed at the same time are shown in Figure 12. The profile for the bar at the mouth of a major distributary (Kumak Channel) has been projected to the surface to demonstrate an average annual surface temperature of -8°C to -9°C. This trend is also projected downwards to the 0°C level, giving an estimated permafrost thickness of about 75 m. This near-linear trend suggests that freezing has been uninterrupted since this location first became exposed to the estimated surface temperature. It provides a reference against which temperature profiles beneath other bars may be compared.

In Figure 12, the profile beneath a recently accreted portion of a bar at the mouth of a medium-sized distributary (Harry Channel, Fig. 11) shows much warmer temperatures. The approach to 0°C at 10 m depth suggests that thawing to this depth has occurred in the recent past. If this site was subaerially exposed for an interval of time similar to the Kumak Channel site, then the present temperature profile indicates an interval of recent submergence that is probably due to erosion of the bar. The bar has subsequently re-emerged, presumably due to accretion, and refreezing is underway.

The time elapsed since the beginning of the thawing episode for the Harry Channel bar can be estimated by using a mathematical model of temperature change and thawing in response to a sudden warming. In Figure 12 the theoretical result of warming from a subaerial -8°C or -9°C to a submerged 4°C is shown for the intervals of 10, 20, and 30 years after submergence of the bar. The profile for 30 years most closely matches the measured temperatures at the Harry Channel site. If the Harry Channel site originally had the same distribution of ground temperature as the Kumak Channel site, then the Harry Channel site appears to have been eroded about 30 years before it began to refreeze. During those 30 years, ground temperatures warmed from values similar to those at the Kumak Channel bar to almost the present values at Harry Channel. This process was accompanied by thawing, which extends to about 10 m. The bar outlines in Figure 13 suggest that re-emergence occurred after 1972, at which time refreezing would have begun. Refreezing has just been completed, as indicated by the continuously recorded subzero (degrees Celsius) temperatures (dashed envelope of temperature variation in Fig. 10) over the previously thawed depth interval. The original bar appears to have eroded, perhaps by a storm that occurred about 30 years prior to 1972, i.e. in the early 1940s.

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Streamflow in the Mackenzie valley

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Abstract: Rivers in the Mackenzie valley generally exhibit a spring-snowmelt dominated mean monthly flow pattern classified as a subarctic, nival flow regime. Along the Mackenzie River, ice jamming during spring breakup is an important process since it can obstruct flow locally and elevate river stage to flood levels. Flow along the large tributary rivers draining the Mackenzie Mountains to the west of the Mackenzie River can rise and fall rapidly in response to rain storms and diurnal snow melt. Global climatic change may result in an increase in the magnitude and frequency of extreme floods, in common with mid-latitude rivers. Some hydraulic modelling suggests that precipitation, and therefore runoff, will increase, but other models indicate that the increased precipitation will be offset by greater evapotranspiration. Considerably more immediate and drastic changes to the flow regime of a given river could arise from dam construction for hydroelectric generation.

Résumé : Les rivières de la vallée du Mackenzie affichent généralement un débit mensuel moyen qui montre l'effet prédominant de la fonte nivale printanière, définissant un régime d'écoulement que l'on qualifie de nival subarctique. Tout le long du fleuve Mackenzie, la formation d'embâcles de glace durant la débâcle printanière est un phénomène important puisque l'obstruction de l'écoulement local par les glaces produit une élévation des eaux du fleuve jusqu'aux niveaux d'inondation. Le débit des principaux affluents qui drainent les monts Mackenzie à l'ouest du fleuve Mackenzie peut fluctuer rapidement suite à des pluies torrentielles ou en raison de la fonte diurne de la neige. Le réchauffement climatique à l'échelle planétaire se traduit par des crues extrêmes plus intenses et plus fréquentes, comme sous les latitudes moyennes. Certains modèles hydrauliques indiquent que les précipitations et, par conséquent, le ruissellement, augmenteront, alors que d'autres modèles laissent entendre qu'une augmentation des précipitations sera compensée par un accroissement de l'évapotranspiration. La construction de barrages pour la production d'électricité pourrait causer des changements considérablement plus rapides et importants au régime d'écoulement d'un cours d'eau donné.

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INTRODUCTION

The Mackenzie River network represents an integral component of the geomorphic landscape in the Mackenzie valley and functions as an important transportation network for communities and economic activities in the region. The purpose of this paper is to review the general streamflow characteristics of Mackenzie River and its major tributaries and to discuss the possible implications on streamflow of global climatic change and flow regulation by hydroelectric dam construction.

DATA

Streamflow in the Mackenzie valley is measured routinely by the Water Survey of Canada (Water Survey of Canada, 1992). For such a large area, the density of stations is very low compared to southern Canada, and many large rivers are ungauged (e.g. Hare Indian River, North Nahanni River). Except for several stations along Mackenzie River and its most important Mackenzie valley tributary, Liard River, streamflow records are of relatively short duration, beginning in the early 1970s or later (Table 1). For some stations, only partial records exist for the initial several years. Many of these stations were established in response to proposed pipelines, highways, and hydroelectric power projects along the Mackenzie valley, most of which were not built.

FLOW REGIME

Generally, rivers in the Mackenzie valley exhibit a spring-snowmelt-dominated mean monthly flow pattern classified as a subarctic, nival flow regime (*see* Church, 1974, 1977). The maximum mean monthly discharge occurs in the May to July period, mainly due to snowmelt. As the snow pack wanes, flow gradually becomes dependent upon base flow and discharge generally decreases slowly through the summer months. This decrease in flow continues throughout the fall and into the winter as ground temperatures decline and freeze-up begins. Storm-induced flows cause periodic increases in discharge that may significantly exceed the

Table 1. Background information about steamflow recording stations.

Station	Station number	Record	Location (lat./long.)	Drainage area (km^2)	River stage[a] (m)		
					At mean annual discharge (m^3/s)	At maximum mean monthly discharge (m^3/s)	At maximum historic daily discharge (m^3/s)
Mackenzie River at Arctic Red River	10CL014	1972–1992	67°27'30"N 133°44'41"W	1 660 000	3.45 (9140)	6.48 (21 400)	8.30 (35 000)
Mackenzie River at Norman Wells	10KA001	1943–1992	65°16'21"N 121°21'25"W	1 570 000	3.94 (7760)	6.13 (17 300)	8.50 (33 300)
Mackenzie River at Fort Simpson	10GC001	1938–1992	61°52'07"N 117°29'48"W	1 270 000	4.83 (6770)	7.03 (14 200)	9.62 (23 500)
Mackenzie River near Fort Providence	10FB001	1961–1978	61°15'04"N 117°29'48"W	971 000	151.21 (4280)	152.28 (7160)	152.83 (8830)
Liard River near the mouth	10ED002	1972–1992	61°44'50"N 121°13'25"W	277 000	2.70 (2490)	5.56 (7600)	8.58 (16 100)
Great Bear River at outlet of Great Bear Lake	10JC003	1961–1992	65°08'05"N 123° 31'06"W	145 000	1.89 (534)	2.05 (590)	2.19 (644)[b]
Peel River above Fort MacPherson	10MC002	1969–1992	67°14'10"N 134°54'28"W	70 600	5.01 (694)	7.67 (2540)	12.24 (8800)
Willowlake River above Metahdali Creek	10GB006	1975–1992	62°39'01"N 122°53'51"W	20 200	4.58 (71)	6.12 (335)	9.50 (1910)
Arctic Red River near the mouth	10LA002	1968–1992	67°47'19"N 133°04'46"W	18 600	2.29 (158)	3.57 (532)	7.94 (3000)
Redstone River 63 km above mouth	10HB005	1974–1992	63°55'32"N 125°18'02"W	15 400	2.69 (181)	3.37 (495)	5.30 (3030)[b]
Mountain River below Cambrian Creek	10KC001	1978–1992	65°13'38"N 128°34'00"W	11 100	3.86 (123)	4.70 (403)	5.50 (1400)
Root River near the mouth	10GA001	1974–1992	62°28'34"N 123°25'49"W	9820	2.70 (95)	3.28 (235)	9.763 (5730)
Blackwater River at outlet of Blackwater Lake	10HC006	1986–1992	63°54'06"N 123°19'22"W	7850	1.52 (47)	5.19 (181)	6.63 (573)
Ramparts River near Fort Good Hope	10KD004	1985–1992	66°07'00"N 129°16'16"W	7410	1.88 (43)	2.84 (167)	5.00 (662)[b]
Carcajou River below Imperial Creek	10KB001	1976–1992	65°17'52"N 127°41'05"W	7400	4.17 (74)	5.03 (220)	8.90 (1900)[b]

[a] River stages were extrapolated from Water Survey of Canada discharge-stage graphs and are relative to arbitrary datum points at the location of the gauging stations i.e. a stage of 0 m does not necessarily correspond to a discharge of 0 m^3/s.

[b] Maximum stage and discharge on the available discharge-stage graph. Higher discharges have been reported for these rivers, estimated by extrapolating the existing curves.

highest mean monthly flow, particularly for rivers draining the Mackenzie Mountains to the west of Mackenzie River. From January to March, discharge is very low because precipitation occurs as snow and base flow is severely limited by the deeply frozen land surface. The onset of snow melt in April causes the discharge cycle to repeat itself.

The subarctic, nival regime is readily apparent on the hydrographs of Liard, Root, Redstone, Carcajou, Mountain, Ramparts, Arctic Red, Peel, and Willowlake rivers, and Mackenzie River at Fort Simpson, Norman Wells and Tsiigehtchic (formerly Arctic Red River) (Fig. 1). These hydrographs depict average mean monthly discharge, but there is considerable variation in mean monthly flow from year to year, particularly between April and November, as illustrated by the plus and minus two standard deviations of mean monthly flows plotted on the selected hydrographs of Figure 2. Daily streamflow during these months, however, may vary considerably, and even significantly depart from mean month flow conditions (Fig. 2) because of fluctuations in short-term weather conditions (e.g. periods of warming and cooling, rainstorms).

The subarctic, nival flow regime is attenuated where large lakes are present within a watershed (e.g. Mackenzie River at Fort Providence, Great Bear River). Large lakes function as natural reservoirs where water is stored and released gradually, thus minimizing extreme discharges and decreasing the variation between maximum and minimum mean monthly flows. The degree of flow regime dampening is a function of the lake basin size, the size of the lake relative to the drainage basin, and proximity of the stage recorder to the lake outlet. The attenuation of runoff is very pronounced for both daily and mean monthly flows for the Great Bear River station at the outlet of Great Bear Lake, as the surface area of the lake represents 21% of the drainage basin (Fig. 2c).

Comparison of the four hydrographs for Mackenzie River (Fig. 1) reveals an obvious change in the flow regime downstream along the Mackenzie valley. At Fort Providence, Mackenzie River has an attenuated discharge regime reflecting the proximity of the station to the outlet of Great Slave Lake. The late spring, summer, and fall mean monthly flows increase significantly downstream between the successive stations because of the addition of runoff from the tributary basins. Overall, the subarctic, nival regime of Mackenzie River becomes more developed downstream. During the winter, the downstream increase in mean monthly flow is relatively insignificant because of the very low winter tributary flows being added to the Great Slave Lake outflow. The most important supplement to Mackenzie River discharge comes from the largest tributary, Liard River. Flow from Liard River produces a marked increase in the summer and fall mean monthly flows of the lower Mackenzie River; the Fort Simpson hydrograph essentially consists of the Liard River and Fort Providence hydrographs superimposed upon one another.

ICE JAMS

Ice jamming along the river channels during the annual spring breakup is an important process in the Mackenzie River system. Major ice jamming in Mackenzie River often occurs as snowmelt runoff from the tributaries is being routed into the trunk stream at the same time as the Mackenzie River ice cover is breaking up. Large ice jams can severely obstruct river flow locally causing a major rise in river stage. A reconstruction of flood events at selected communities along Mackenzie, Liard, and Peel rivers by Kriwoken (1983) found that most large floods were associated with high backwater conditions generated by large ice jams situated downstream. Determining the levels and frequencies of such floods must often rely upon geomorphic evidence along the river banks or historical records rather than flow-based flood frequency relationships, because stage-discharge curves are not applicable under backwater conditions generated from ice jams or water-level data is either unavailable or available for only a short duration.

The critical role of Liard River streamflow in initiating the spring breakup of Mackenzie River is particularly noteworthy. In most years, the winter ice cover of Mackenzie River downstream of the Mackenzie–Liard confluence breaks up before the section of river upstream (MacKay and Mackay, 1973). This breakup pattern is triggered by a rise in stage of Mackenzie River resulting from the influx of the Liard River spring flow (Prowse, 1984). The addition of the Liard spring flow exerts great pressure on the Mackenzie River ice cover resulting in the lifting and fracturing of the ice mass (MacKay and Mackay, 1973). The actual breakup commonly is caused by the physical jamming and push of broken-up Liard ice into the Mackenzie River ice cover. In the section of Mackenzie River upstream, the ice cover is not subject to a significant rise in stage due to the damping effects on discharge caused by Great Slave Lake and thus breakups occur later than below the Mackenzie–Liard confluence.

MACKENZIE RIVER TRIBUTARIES

Many of the Mackenzie tributaries occupy large river basins that appear relatively small when compared to the continental scale of the Mackenzie River system. Rivers draining the Mackenzie Mountains to the west of Mackenzie River have 'flashy' discharge regimes, because streamflow can rise and fall rapidly in response to rainstorms and diurnal snowmelt. This characteristic is not apparent from the mean monthly discharge patterns depicted in Figure 1, but can be seen in the 1991 mean daily hydrographs of Redstone and Willowlake rivers (Fig. 2a, 2d.). Rivers draining the lowlands and plains east of Mackenzie River have a less responsive discharge regime because of the presence of lakes and extensive bogs, which attenuate runoff from both snowmelt and rainstorm events. Also, the more subdued plain and lowland topography east of Mackenzie River does not produce widespread intense rainfall, as generated the major orographic uplift of moist air masses over the Mackenzie Mountains to the west.

All of the tributaries, except for those whose flow regime is conditioned by large lakes (e.g. Great Bear River and, to a lesser extent, Blackwater River), experience large periodic increases in discharge, due to severe summer rainstorms (see Mackay et al., 1973; Jasper and Kerr, 1992). Because such events only last for a few days and may not reoccur for up to 20–30 years, the magnitude of extreme discharges are not always apparent in historic maximum daily discharge and

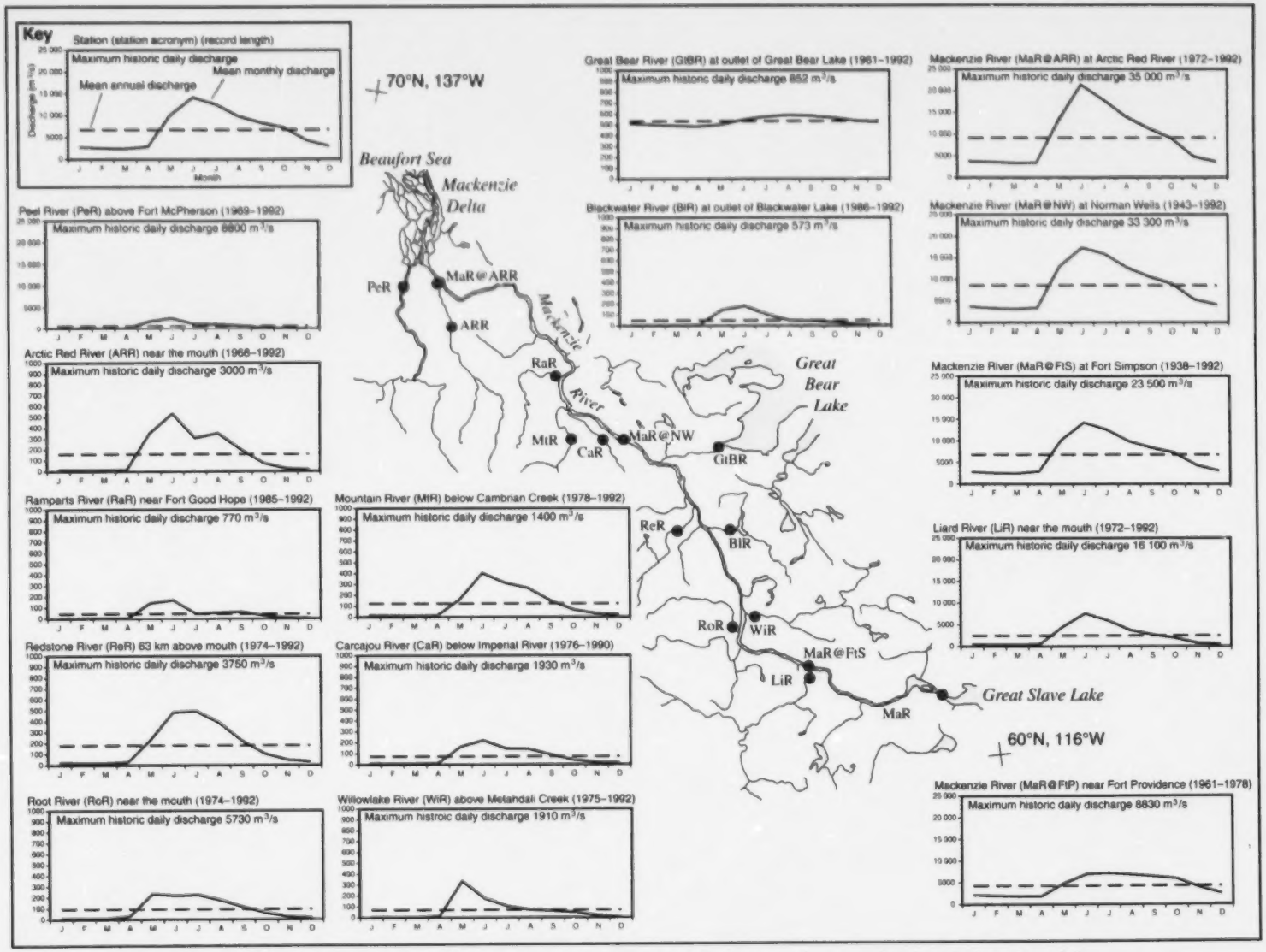
Key
Station (station acronym) (record length)
Maximum historic daily discharge
Mean monthly discharge
Mean annual discharge
Discharge (m³/s)
Month
70°N, 137°W
Beaufort Sea
Mackenzie Delta
PeR
MaR@ARR
ARR
Mackenzie River
RaR
MtR
CaR
MaR@NW
GtBR
Great Bear Lake
ReR
BlR
RoR
WiR
MaR@FtS
LiR
MaR
Great Slave Lake
60°N, 116°W
Great Bear River (GtBR) at outlet of Great Bear Lake (1961–1992)
Maximum historic daily discharge 852 m³/s
Blackwater River (BlR) at outlet of Blackwater Lake (1986–1992)
Maximum historic daily discharge 573 m³/s
Peel River (PeR) above Fort McPherson (1969–1992)
Maximum historic daily discharge 8800 m³/s
Arctic Red River (ARR) near the mouth (1968–1992)
Maximum historic daily discharge 3000 m³/s
Ramparts River (RaR) near Fort Good Hope (1985–1992)
Maximum historic daily discharge 770 m³/s
Mountain River (MtR) below Cambrian Creek (1978–1992)
Maximum historic daily discharge 1400 m³/s
Redstone River (ReR) 63 km above mouth (1974–1992)
Maximum historic daily discharge 3750 m³/s
Carcajou River (CaR) below Imperial River (1976–1990)
Maximum historic daily discharge 1930 m³/s
Root River (RoR) near the mouth (1974–1992)
Maximum historic daily discharge 5730 m³/s
Willowlake River (WiR) above Metahdali Creek (1975–1992)
Maximum histroic daily discharge 1910 m³/s
Mackenzie River (MaR@ARR) at Arctic Red River (1972–1992)
Maximum historic daily discharge 35 000 m³/s
Mackenzie River (MaR@NW) at Norman Wells (1943–1992)
Maximum historic daily discharge 33 300 m³/s
Mackenzie River (MaR@FtS) at Fort Simpson (1938–1992)
Maximum historic daily discharge 23 500 m³/s
Liard River (LiR) near the mouth (1972–1992)
Maximum historic daily discharge 16 100 m³/s
Mackenzie River (MaR@FtP) near Fort Providence (1961–1978)
Maximum historic daily discharge 8830 m³/s

Figure 1. *Hydrographs for the Mackenzie and eleven major tributaries and a map showing the location of the streamflow gauging stations. Each hydrograph shows the mean monthly flow (m^3/s) and the maximum recorded daily flow. Note that hydrographs are plotted at two different scales, 0–1000 m^3/s and 0–25 000 m^3/s.*

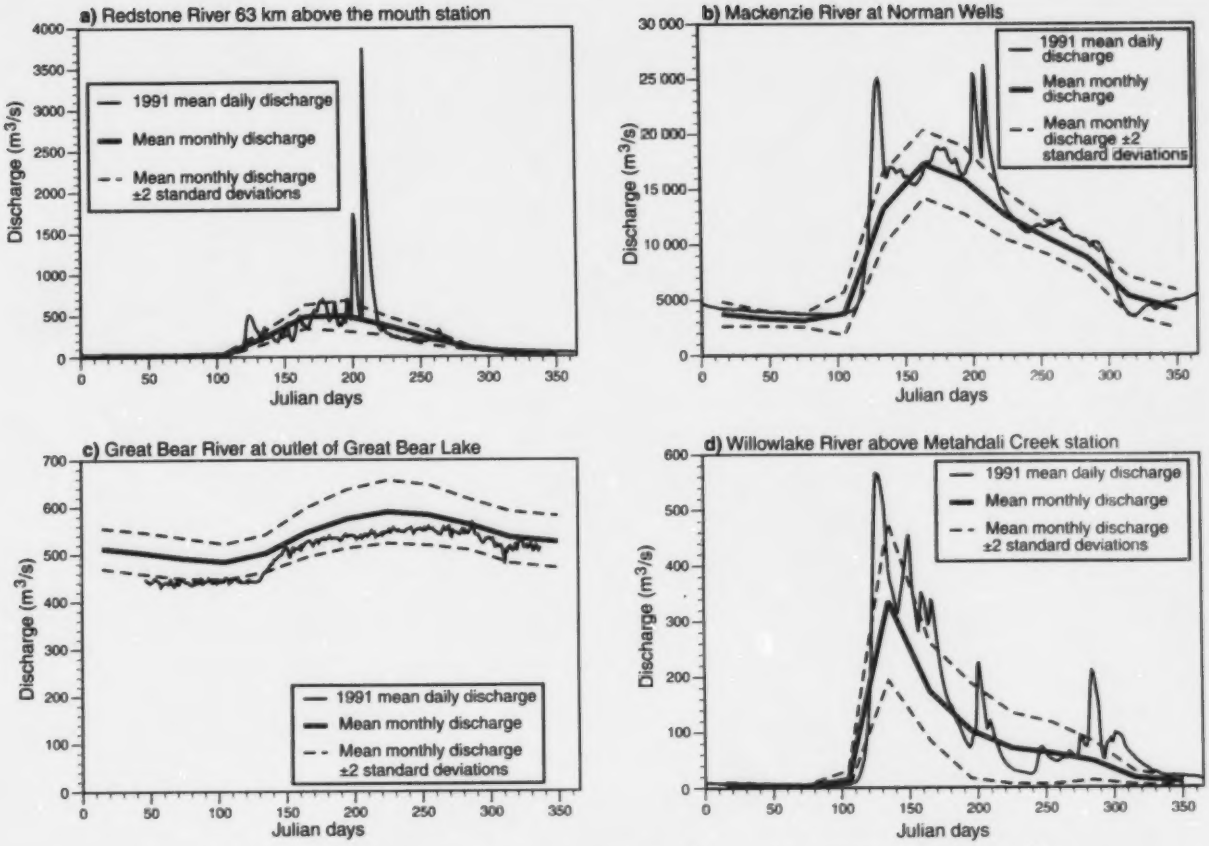
a) Redstone River 63 km above the mouth station
1991 mean daily discharge
Mean monthly discharge
Mean monthly discharge ±2 standard deviations
Discharge (m3/s)
Julian days
b) Mackenzie River at Norman Wells
1991 mean daily discharge
Mean monthly discharge
Mean monthly discharge ±2 standard deviations
Discharge (m3/s)
Julian days
c) Great Bear River at outlet of Great Bear Lake
1991 mean daily discharge
Mean monthly discharge
Mean monthly discharge ±2 standard deviations
Discharge (m3/s)
Julian days
d) Willowlake River above Metahdali Creek station
1991 mean daily discharge
Mean monthly discharge
Mean monthly discharge ±2 standard deviations
Discharge (m3/s)
Julian days

Figure 2. *Selected hydrographs showing the 1991 mean daily discharge and the mean month discharge (shown to two standard deviations) based upon the entire streamflow record, for the four stations.*

mean monthly discharge records for stations of relatively short duration (compare the daily and mean monthly hydrograph of Redstone River in Fig. 2a). For example, an early July 1988 storm generated a mean daily discharge along Root River that was a full order-of-magnitude larger than the highest mean monthly discharge. Maximum daily discharges for the period of record have therefore been added to the hydrographs in Figure 1 to illustrate the upper limit of flows recorded at each station. The short (and in some cases incomplete) discharge record for many stations may therefore be insufficient to have recorded extreme events. Thus, the maximum daily discharge for many stations is unrepresentative of the extreme flows expected over longer time periods.

CLIMATE CHANGE

The effects of global climatic change upon the streamflow of the Mackenzie valley river systems are not known with certainty, and only recently have been the subject of investigation. Hydrological modelling at the global scale suggests that there will be an increase in runoff for rivers at the high latitudes of North America, in response to an increase in precipitation (Miller and Russell, 1992). More locally, preliminary information from Soulis et al. (1994) for a number of Mackenzie basin watersheds also indicates that precipitation will increase, which suggests an increase in runoff. Some of the modelling scenarios of Soulis et al. (1994), however, indicates that this increased precipitation will be offset by greater evapotranspiration making the landscape drier than present. These contradictory results need to be clarified, but perhaps suggest that the net effects upon average streamflow may be minimal. Another possible effect of climate warming upon streamflow will be an earlier onset and longer period of spring melt. Although not related to streamflow, additional factors that could be affected by climate warming include an earlier spring breakup and later fall freeze-up, and reduced winter ice cover which may lead to less severe ice jamming during the spring breakup.

Global climatic change may also have a major impact upon the frequency of flood events, which have been shown to increase in some regions of North America (Knox, 1993). Flood events are very important to the discharge regimes and geomorphology of many Mackenzie valley rivers and can induce considerable change in channel morphology along some river reaches (*see* Brooks, 2000). The effects of an increased flooding frequency may be most pronounced along braided rivers, which can experience major channel change during these events.

EFFECTS OF DAMS

It should be noted, however, that considerably more immediate and drastic changes to the flow regime of a given watershed than those caused by global change could arise from dam construction for hydroelectric generation. Many of the large Mackenzie valley tributaries were assessed for hydroelectric potential in the early 1970s, but this work has never proceeded beyond the initial evaluation (E.B. Owen, unpub. report, 1973). Beginning in 1967, however, streamflow of Mackenzie River has been partially regulated by operations of Bennett Dam in the upper Peace River basin, northeastern British Columbia. As exemplified by the gauging station at Arctic Red River, the total annual inflow of Mackenzie River is unchanged, but the average flows during the winter months have increased 20–30% while the summer and fall flows have diminished up to 5–10% (Wiens, 1991). The regulatory effect of Bennett Dam upon the flow of Mackenzie River at Tsiigehtchic is relatively insignificant, because the dam is situated well upstream of the natural regulation by Great Slave Lake, and because of the addition of flow from numerous unregulated tributary basins located between Great Slave Lake and Tsiigehtchic.

Particularly significant, however, would be the effects of hydroelectric development in the Liard Basin, although this seems unlikely under present economic conditions. Liard River is the large tributary of Mackenzie River whose flow is not regulated by a large lake. As is well documented, Liard River is an important sediment source and has a critical role in initiating the spring breakup of the Mackenzie River downstream of the Laird–Mackenzie confluence. If Liard River becomes regulated by a dam, significant effects include a reduced Liard freshet, reduced sediment supply affecting areas downstream as far as the delta, an increase in the daily fluctuation of Liard River discharge, and a delay in the spring breakup of Mackenzie River (Hardy Associates (1978) Ltd., 1982).

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I thank Pat Wood (Fort Simpson, Water Survey of Canada) and Moe Hansen (Inuvik, Water Survey of Canada) for numerous discussions about Mackenzie River and its tributaries.

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Channel changes along the lower reaches of major Mackenzie River tributaries

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Brooks, G.R., 2000: Channel changes along the lower reaches of major Mackenzie River tributaries; in *The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 159–166.*

Abstract: Alluvial rivers undergo lateral channel change through progressive lateral migration at bends or through channel avulsions. The floodplains of rivers experiencing lateral channel change contain geomorphic features indicative of past migration. Based on aerial photography spanning from the late 1940s–early 1950s to the 1970s–1980s, insignificant lateral channel change occurred along the lowest reaches of the Root, Willowlake, Blackwater, Great Bear, Hume, Ramparts, Hare Indian, Ontaratue, and Arctic Red rivers; all of which have a meandering or meandering-straight planform. In contrast, the lowest reaches of the North Nahanni, Dahadinni, Redstone, Keele, Mountain and Carcajou rivers have all experienced significant lateral channel change, ranging from up to 3 m/year to 11 m/year. An increase in the magnitude and frequency of extreme flows arising from climate warming may cause an overall increase in the rate of bank erosion and net widening of the channel along those river reaches presently experiencing active lateral migration.

Résumé : Les cours d'eau de plaine alluviale voient leur chenal se déplacer latéralement à cause de la migration latérale progressive des méandres ou du détournement du chenal par avulsion. Les plaines d'inondation des cours d'eau se déplaçant latéralement contiennent des éléments géomorphologiques qui témoignent d'une migration antérieure. L'analyse de photographies aériennes prises durant la période allant de la fin des années 1940 et du début des années 1950 aux années 1970 et 1980 révèle qu'il y a eu un déplacement latéral non significatif du lit dans les tronçons inférieurs des rivières Root, Willowlake, Blackwater, Great Bear, Hume, Ramparts, Hare Indian, Ontaratue et Arctic Red; toutes ont une forme en plan méandrique ou méandrique-rectiligne. Par contre, les tronçons inférieurs des rivières Nahanni Nord, Dahadinni, Redstone, Keele, Mountain et Carcajou ont subi un changement latéral important de leur lit qui s'est produit à un rythme d'environ 3 à 11 m/a. L'augmentation de l'intensité et de la fréquence des crues extrêmes causées par le réchauffement climatique pourrait causer une accélération globale de l'érosion des berges et un élargissement net des chenaux qui subissent actuellement une migration latérale.

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INTRODUCTION

Alluvial rivers flowing within a valley bottom containing a floodplain inherently undergo channel change. Over time this can result in lateral migration of the river within the valley bottom, and the widening and narrowing of the river channel. The specific type and relative magnitude of these processes are controlled by local factors and thus can vary considerably from river to river and between different reaches of the same river. Large sections of the major tributaries to the Mackenzie River are alluvial rivers. The purpose of this paper is to briefly review lateral channel change and channel widening processes, summarize channel migration and channel widening occurring along the lower reaches of major Mackenzie River tributaries, and generally discuss the possible implications of climate change on channel change processes.

LATERAL CHANNEL CHANGE

Alluvial rivers undergo lateral channel change through progressive lateral migration and/or avulsions. Progressive lateral migration occurs at river bends through concomitant erosion along the concave (outer) banks and deposition along the convex (inner) banks. This process occurs along both meandering and braided planforms, but is most often associated with the meandering rivers. Along a meandering river, progressive lateral migration over time results in the outward expansion and downvalley procession of the river meanders (Fig. 1). The rate of lateral migration along a river reach is dependent upon a number of local factors, specifically: stream power, planform geometry, sediment transport, bank height, and erosional resistance of the concave banks (Hickin and Nanson, 1984).

An avulsion occurs when a river creates a new channel or abandons an old one. There are several types of avulsions: 1) the meander cutoff process which results in a new section of channel being formed across the neck of the meander being cutoff (Fig. 2a); 2) the creation or abandonment of a channel across the floodplain or vegetated island along a braided channel network (Fig. 2b); and 3) the total abandonment and relocation of the existing channel (Fig. 2c). Avulsions are most likely to occur during periods of high discharge when a river overtops its banks and begins to flow within abandoned channels, depressions, or irregularities on the floodplain. A new or reactivated channel may then be eroded into the floodplain which eventually captures all or part of the river flow.

The floodplains of rivers experiencing lateral channel change contain geomorphic features indicative of past lateral instability. These features are readily apparent on aerial photographs and include abandoned channels, channel scars,

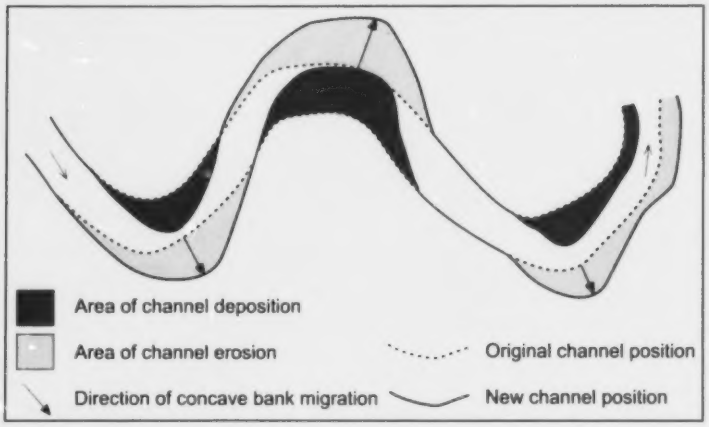

Figure 1.

The process of progressive lateral channel migration along a meandering river through the concomitant erosion of the concave bank and deposition along the convex bank.

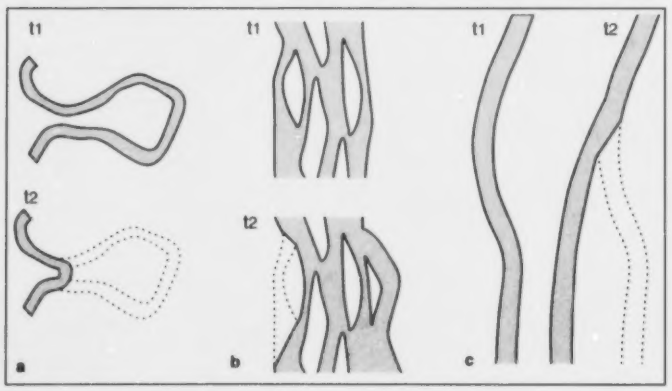

Figure 2.

The three basic types of avulsions: ***a)*** *the meander cutoff process;* ***b)*** *the creation and/or abandonment of channels within a braided channel network; and* ***c)*** *the total abandonment and relocation of an existing channel. The channel positions before and after the avulsions are indicated by the times* t_1 *and* t_2*, respectively.*

ox-bow lakes, lateral successions of vegetation communities which are progressively older beginning from the convex banks, and curvilinear tracts of different types and age of vegetation communities (Fig. 3a, 3b).

DYNAMIC CHARACTER OF CHANNEL WIDTH

Channel change processes operate most intensely during periods of high flow (spring runoff or storm events), after a threshold discharge has been exceeded which allows a river to erode the sediments along the base of the concave bank. In some instances, this erosion may be removing sediments that have wasted from the bank face during periods of low flow and which have accumulated along the edge of the channel, rather than actually eroding the intact bank face.

Along a given river reach, concave bank erosion varies considerably from year to year because of variation in the magnitude of the annual high flow. Research on mid-latitude meandering rivers indicates that major concave bank erosion can occur during infrequent, very high-magnitude discharge events (e.g. Pizzuto, 1994) causing significant lateral displacement of the channel. This can also result in major widening of the channel when the relative amount of concave bank erosion greatly exceeds the deposition along the convex bank. Such infrequent, high-magnitude events commonly are followed by a period of successive years having lower, more frequent, high discharges during which the relative amount of aggradation along the convex bank exceeds the erosion occurring along the concave bank, thus causing the channel width to slowly narrow. After a period of time (years or decades), the recurrence of a very high-discharge event may again cause major erosion of the concave bank and substantially widen the channel. The episodic occurrence of major concave bank erosion followed by periods of net aggradation allow the channel cross-section to respond dynamically to discharge conditions over time through widening or narrowing of the channel. Overall, at any give time, channel morphology of a river will reflect the recent discharge history.

Braided rivers generally are also sensitive to the occurrence of infrequent, very high-discharge events. Such events may cause significant progressive erosion of the concave banks of channels both within and along the margin of the braided network resulting in erosion of floodplain, vegetated islands, and stabilized bars. This may be accompanied by avulsions that increase the number of channels. The net result of these processes is the lateral migration of channel(s) and the net increase in channel width. The occurrence of infrequent, very high-magnitude discharges have been documented to cause dramatic widening of some braided river reaches. Again, the infrequent, high-magnitude events commonly are followed by a period of successive years having

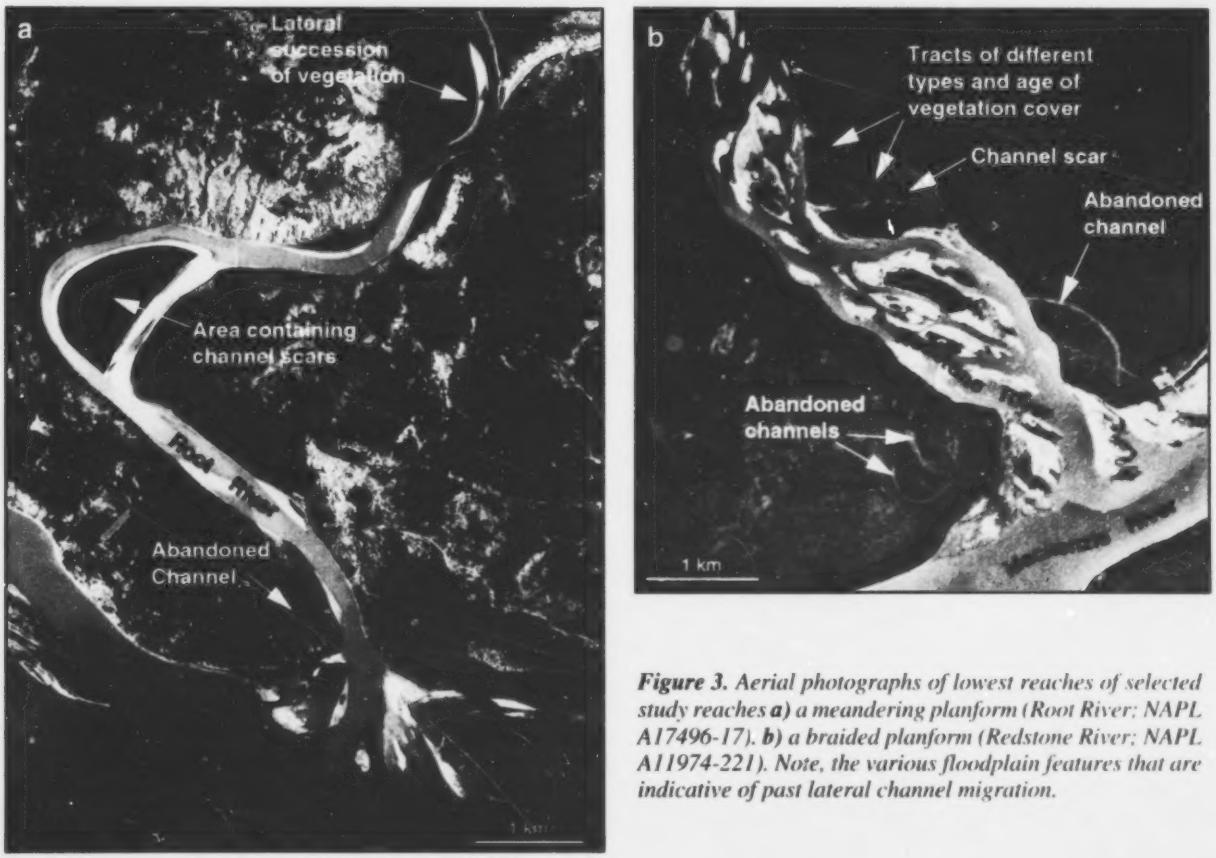

Figure 3. *Aerial photographs of lowest reaches of selected study reaches* ***a)*** *a meandering planform (Root River; NAPL A17496-17).* ***b)*** *a braided planform (Redstone River; NAPL A11974-221). Note, the various floodplain features that are indicative of past lateral channel migration.*

lower, more frequent high discharges, in which net aggradation occurs in the channel(s) causing the river width to gradually narrow. This occurs because of reduced concave bank erosion in combination with bar accretion, channel abandonment, and revegetation of the inactive bar surfaces. After a sufficient period of channel 'recovery', the recurrence of another infrequent, high-magnitude discharge will repeat the widening-narrowing cycle.

Recognition of the cyclic nature of channel migration and widening is important in determining rates of lateral channel migration (Nanson and Hickin, 1983). It is critical that channel change be assessed over a sufficient period of years so that a very wide range of peak flows and storm events have occurred within the intervening period to gain a proper understanding of local channel change characteristics and to allow a representative average of channel migration to be determined. Representative average rates of channel change can be determined from historic aerial photographs or from dendrochronology of the forest on the floodplain behind the convex bank (*see* Hickin, 1988).

CHANNEL CHANGE ALONG THE MACKENZIE VALLEY

Overall, few channel change studies have been conducted on the Mackenzie valley river systems. G. Brooks (unpub. data, 1994; Brooks, 1996) mapped lateral channel migration along the lowest 4–10.5 km of 15 major tributaries entering Mackenzie River between Fort Simpson and the Mackenzie Delta (Fig. 4) using aerial photography beginning in the late 1940s or early 1950s and extending into the 1970s or the 1980s, depending upon availability. Rates of lateral channel change were based upon the amount of erosion occurring along the steep concave banks which on aerial photographs

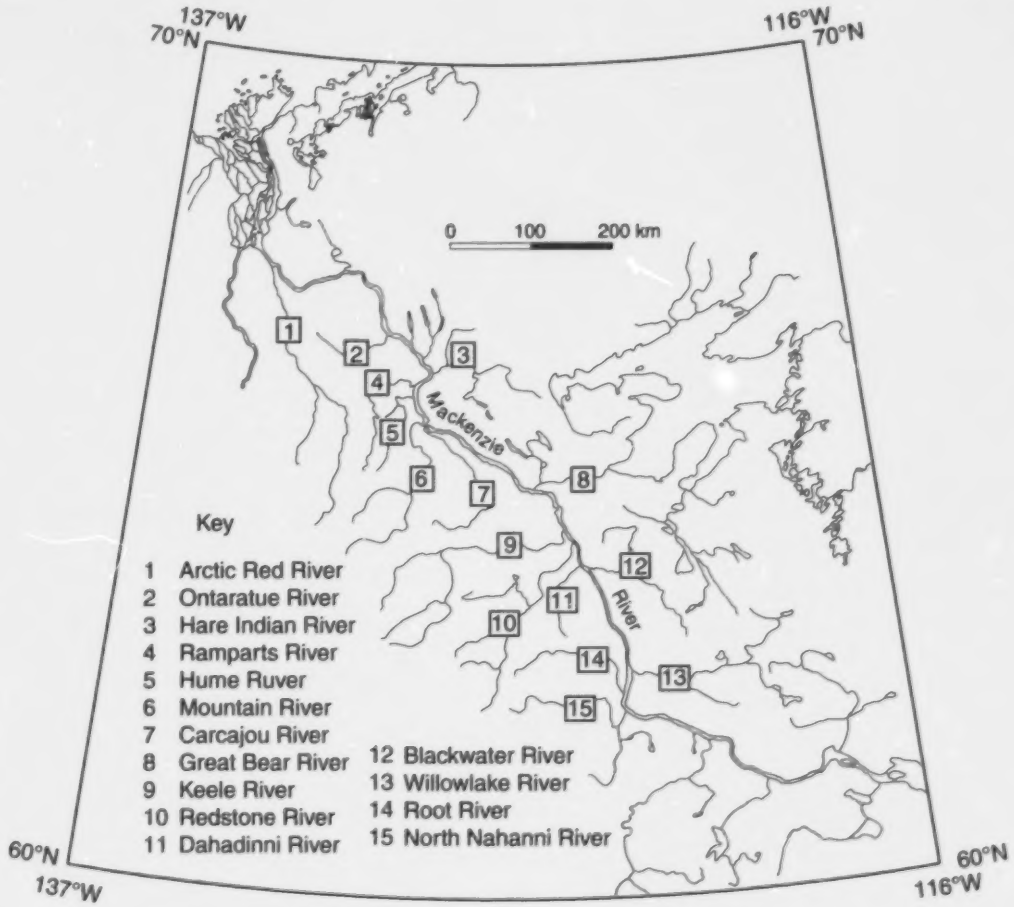

Figure 4. *Map showing the location of 15 major tributaries of the Mackenzie River. Channel change data of the lowest 4–10.5 km of each tributary is summarized in Table 1.*

are well defined by a sharp break in the vegetation cover at the edge of the river. Rates of lateral accretion were not determined because the margin of the water surface on the river bars is affected considerably by stage differences between sets of aerial photographs and thus is subject to considerable error. Examples of the net historic channel change occurring over the lower reach of a selected meandering and braided study reach are depicted in Figure 5. From Figure 5b, it is obvious that channel change within a braided river can be very complex resulting in modification to both the outer margins of and the islands and bars within the braided network.

Table 1 lists the maximum average rate of lateral channel change occurring along the lowest 4–10.5 km of the 15 tributaries. Bank erosion of 50 m is considered to be the detection level of the methodology used to determine the amount of channel change, thus erosion of less than 50 m is reported as insignificant in Table 1. In Table 1, the rate of maximum erosion along the braided rivers relates to the outer margin of the river and corresponds generally to the concave bank of a major channel meandering back and forth from one side of the braided network to the other. The results in Table 1 represent the average of the maximum rate of erosion occurring within the study reaches over the time span of the aerial photography.

As summarized in Table 1, insignificant lateral channel change has occurred along the lowest reaches of 9 of the 15 tributaries (Root, Willowlake, Blackwater, Great Bear, Hume, Ramparts, Hare Indian, Ontaratue, and Arctic Red rivers). All of these nine rivers have a meandering or meandering-straight planform. Backwater conditions from Mackenzie River exist along seven of these rivers (Root, Willowlake, Hume, Ramparts, Hare Indian, Ontaratue, and Arctic Red rivers) even during relatively moderate discharge levels and this is the major factor limiting lateral channel change. The lower reaches of these rivers are in a sense 'drowned' by the Mackenzie River whereby flow velocity is severely limited causing low stream power. An additional factor limiting lateral channel migration is that some sections of Root, Willowlake, Hume, Hare Indian, and Arctic Red rivers are confined against the relatively high valley sides. Along the remaining two of the nine rivers (Great Bear and Blackwater rivers), the lower reaches are both relatively

Table 1. Summary of channel change processes along lower reaches of selected major Mackenzie River tributaries.

River	Channel planform	Channel change processes	Length of study reach above tributary mouth[a] (km)	Time of airphoto coverage	Maximum lateral channel migration[b] (m)	Maximum rate of lateral channel migration (m/year)
North Nahanni	Braided	Progressive lateral migrations Avulsions Cyclic channel widening	7	1948–1977	200	6.9
Root	Meandering	Progressive lateral migrations Avulsions	8	1947–1977	Insignificant	na
Willowlake	Meandering	Progressive lateral migration	4.5	1947–1972	Insignificant	na
Blackwater	Straight	Progressive lateral migration Avulsions	4	1945–1987	Insignificant	na
Dahadinni	Braided	Progressive lateral migrations Avulsions Cyclic channel widening	6	1945–1981	338	9.4
Redstone	Braided	Progressive lateral migrations Avulsions Cyclic channel widening	6.5	1949–1981	340	10.6
Keele	Braided	Progressive lateral migrations Avulsions Cyclic channel widening	7.5	1949–1981	203	6.3
Great Bear	Meandering	Progressive lateral migration	5.5	1944/45–1983	Insignificant	na
Carcajou	Meandering	Progressive lateral migration Cyclic channel widening (?)	10.5	1950–1981	100	3.2
Mountain	Braided	Progressive lateral migrations Avulsions Cyclic channel widening	8	1950–1981	233	7.5
Hume	Meandering	Progressive lateral migration	5.5	1950–1971	Insignificant	na
Ramparts	Meandering straight	Progressive lateral migration	7	1950–1971	Insignificant	na
Hare Indian	Meandering	Progressive lateral migration	6	1950–1971	Insignificant	na
Ontaratue	Meandering straight	Progressive lateral migration	5.5	1950–1970	Insignificant	na
Arctic Red	Meandering	Progressive lateral migration	7	1950–1980	Insignificant	na

[a] Distance measured along valley axis
[b] The amount of lateral channel migration is averaged over the time interval between sets of air photos. Lateral channel migration of less than 50 m is not considered significant.
na = not applicable

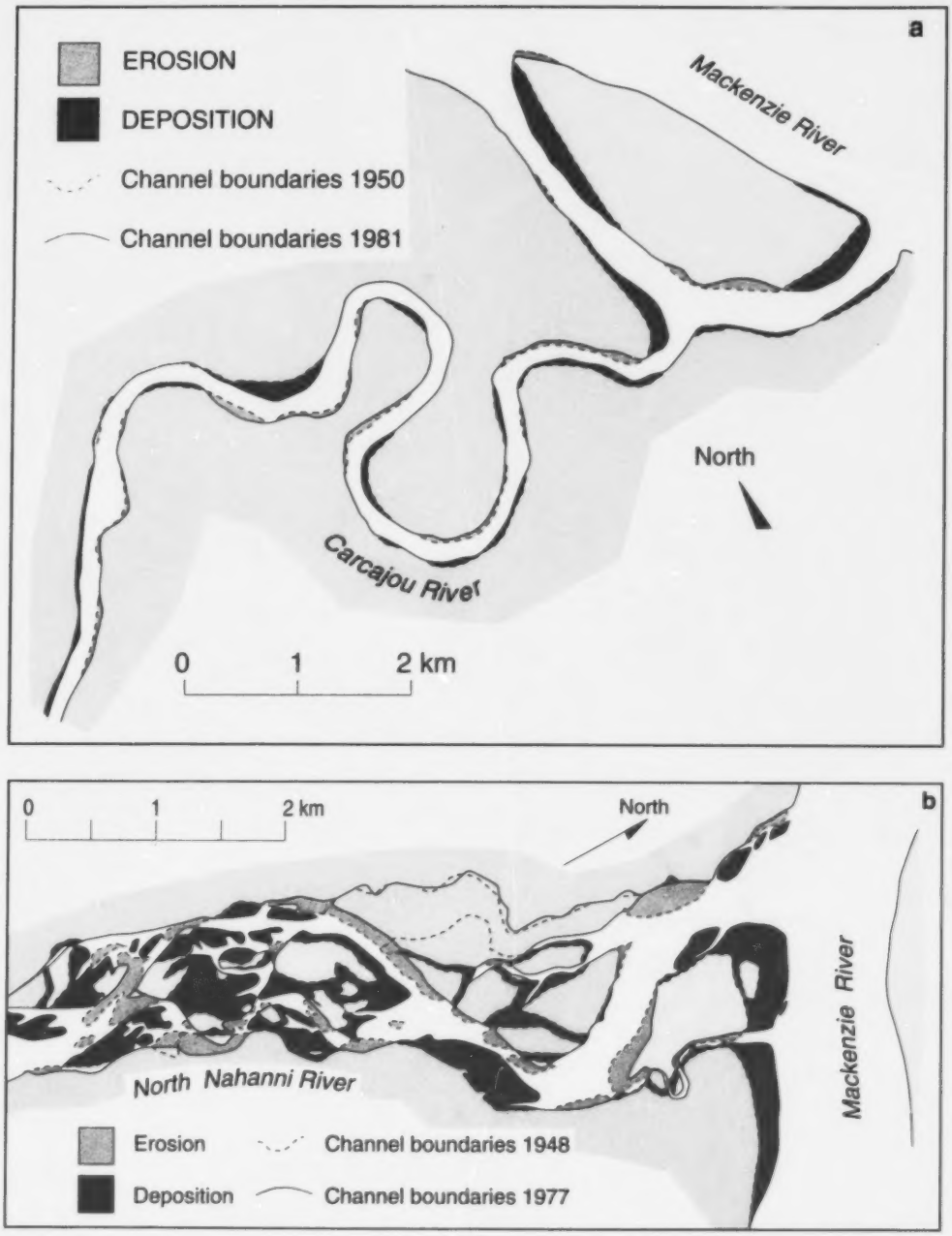

Figure 5. *Net channel migration as determined from historical aerial photographs along:* ***a)*** *the meandering Carcajou River between 1950 and 1981 and* ***b)*** *the braided North Nahanni River between 1948 and 1977. Note that the mouth of Carcajou River is split by an island situated along the edge of the Mackenzie River channel.*

fast-flowing streams. Channel change along Great Bear River is limited by a very regular discharge regime, low sediment load and confinement within a Tertiary bedrock valley. Along the study reach of Blackwater River the channel margins are formed by coarse boulder lag which seems to be resistant to erosion by the river.

The lowest reaches of the five braided rivers (North Nahanni, Dahadinni, Redstone, Keele and Mountain rivers) and the meandering Carcajou River have all experienced significant lateral channel change ranging from about 3–11 m/year. Along these reaches, all five braided rivers are swift flowing, thus, stream power is relatively high, particularly during extreme discharges. Stream power along Carcajou River is lower than the five braided rivers because the flow is slower, but still is sufficiently high to significantly erode the river banks. Significantly, along all six rivers a major proportion of the concave bank erosion is occurring directly into the contemporary floodplains which are relatively low in height and consist of silt, sand, and gravel alluvium that is susceptible to cutbank erosion.

CYCLICAL CHANNEL WIDENING

There has been no study monitoring channel width along any reach of a Mackenzie valley river system. The general effects of high-magnitude flows that have been documented, observed on sequences of aerial photographs and in the field, suggest that cyclic channel widening and narrowing does occur along at least several of the major Mackenzie River tributaries. In 1970, a severe storm centred upon the Mackenzie Mountains caused flooding and major channel changes (including channel widening) along the Redstone, Keele, Mountain, and Arctic Red rivers (MacKay et al., 1973). Early 1970s aerial photography of the Keele, Redstone, and Dahadinni study reaches depict a significantly wider braided channel than that appearing in 1940s aerial photography. Although this widening has occurred sometime in the interval between the 1940 and 1970 aerial photography, it may well have resulted from this flood which preceded the 1970s photography by only a few years. Severe rainstorms in early July 1988 produced major flooding along the upper Mackenzie valley which caused net channel widening along the Keele, Redstone, Dahadinni, and North Nahanni through bank erosion, channel avulsions, and reorganization of the braided channel networks. The geomc.phic effects of this storm were still apparent along these rivers in 1992 in the form of wide areas of inactive channels and bar surfaces that were in the very early stages of revegetation and which probably represent the initial period of channel narrowing.

EFFECTS OF CLIMATE CHANGE

As described in Brooks (2000), hydrological modelling of the effects of global climatic change in the Mackenzie basin area indicates that precipitation would increase resulting in greater stream runoff. Some modelling scenarios, however, suggest that this increased precipitation would be more than offset by greater evapotranspiration, thus making the landscape drier causing a decrease in river runoff. These contradictory results perhaps suggest that the net effect upon average streamflow in the Mackenzie basin may be minimal.

The effects of global climatic change upon the rates of channel change along the lowest reaches of the major Mackenzie River tributaries are not known with certainty. A minor change to mean annual discharge probably will have an insignificant effect upon channel change. In some regions of North America, it is believed that an increase in the frequency of very high-magnitude discharges will be associated with climate change (Knox, 1993). As discussed above, major bank erosion and channel widening can occur along some river reaches during infrequent, very high-magnitude flows. If the frequency of these flows increased, then there could be an overall increase in concave bank erosion and thus an increase in the rate of lateral channel migration. There might also be a net widening of the river channel or braided network because the period between the high-magnitude events would be less thus preventing the same degree of narrowing. Finally, on a smaller scale along the braided rivers, a greater amount of channel shifting within the braided channel networks could occur because of increased erosion and deposition of the river bars.

Increased rates of migration and channel widening could occur along the study reaches of North Nahanni, Dahadinni, Redstone, Keele, Mountain, and Carcajou rivers, all of which currently are experiencing significant amounts of lateral channel change. Of those lower reaches which experienced insignificant channel migration, perhaps only the Blackwater River study reach might be affected if the high-magnitude events have sufficient competence to erode the coarse channel bed. Negligible impacts might be expected along the remaining study reaches. The Great Bear River study reach would continue to experience a very low range of discharge because of the regulating effects of Great Bear Lake whereas backwater conditions from Mackenzie River, which inhibit stream power, would still be present along the lower reaches of Root, Willowlake, Hume, Ramparts, Hare Indian, Ontaratue, and Arctic Red rivers.

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Landslides of the Mackenzie valley and adjacent mountainous and coastal regions

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Aylsworth, J.M., Duk-Rodkin, A., Robertson, T., and Traynor, J.A., 2000: Landslides of the Mackenzie valley and adjacent mountainous and coastal regions; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 167–176.*

Abstract: Approximately 3400 landslides, distinguished by failure class (flow, slide, complex, unclassified) and parent material (bedrock or unconsolidated Quaternary sediment) have been mapped in the Mackenzie valley. Many landslides are related to the degradation of permafrost and ground ice; others are influenced by the control permafrost maintains on ground water movement. Given the prevalence of permafrost and icy sediments in this region, any climate change leading to a rise in ground temperatures and degradation of permafrost will increase the frequency of landslides. The most vulnerable sites include ice-rich, fine-grained sediments on slopes near water bodies or in fire scars, and frozen, coarse-grained sediments overlying clay or clayey till in steep riverbanks where permafrost does not extend to the base of the section.

Résumé : Dans la vallée du Mackenzie, on a répertorié et cartographié quelque 3 400 glissements de terrain que l'on a classés selon le type de rupture (coulée, glissement, complexe, non déterminé) et la nature des matériaux touchés (substratum rocheux ou sédiments quaternaires non consolidés). De nombreux glissements de terrain sont causés par la dégénérescence du pergélisol et de la glace de sol; d'autres dépendent du contrôle qu'exerce le pergélisol sur la circulation des eaux souterraines. Étant donné que le pergélisol et les sédiments gelés dominent dans cette région, tout changement climatique provoquant une hausse des températures du sol et la dégénérescence du pergélisol multiplierait les glissements de terrain. Parmi les sites les plus vulnérables, mentionnons les versants près d'étendues d'eau ou dans les brûlis qui sont découpés dans des sédiments à grain fin riches en glace, ainsi que les berges abruptes formées dans des sédiments à grain grossier gelés reposant sur des argiles ou des tills argileux, où le pergélisol ne descend pas jusqu'à la base de la coupe.

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INTRODUCTION

An inventory of 3400 landslides has been compiled for the Mackenzie valley and adjacent mountainous regions and the Mackenzie Delta and Tuktoyaktuk Peninsula (Aylsworth, 1992; Duk-Rodkin, 1993; Aylsworth and Traynor, in press; Duk-Rodkin and Robertson, in press; Aylsworth et al., in press). This area covers approximately 250 000 km^2 and extends in a 1200 km strip from the Northwest Territories border at 60°N to the Beaufort Sea, encompassing the Mackenzie River and at least the lower reaches of all its tributaries. This corridor also includes the communities of the Mackenzie valley, the Fort Liard, Mackenzie, and Dempster highways and the Norman Wells to Zama oil pipeline (Fig. 1).

Within an area of this size, a tremendous variety of topographical, geological, and thermal conditions are found. Topography ranges from flat or rolling plains to dissected plateaus and rugged mountains. The region is underlain by Cambrian to Tertiary sedimentary bedrock, and, with the exception of steep mountain slopes, most of the valley is covered by Quaternary deposits characterized by till, lacustrine clay, silt and sand, and glaciofluvial and alluvial silt, sand, and gravel (*see* Aylsworth et al., 2000). Permafrost underlies much of the region. Permafrost varies from continuous and thick (>100 m) in the north to sporadic and thin (<5 m) in the extreme south of the valley. Segregated ground ice is common in all fine-grained sediments. Because of the variety of physical conditions in the region, the types of landslides and

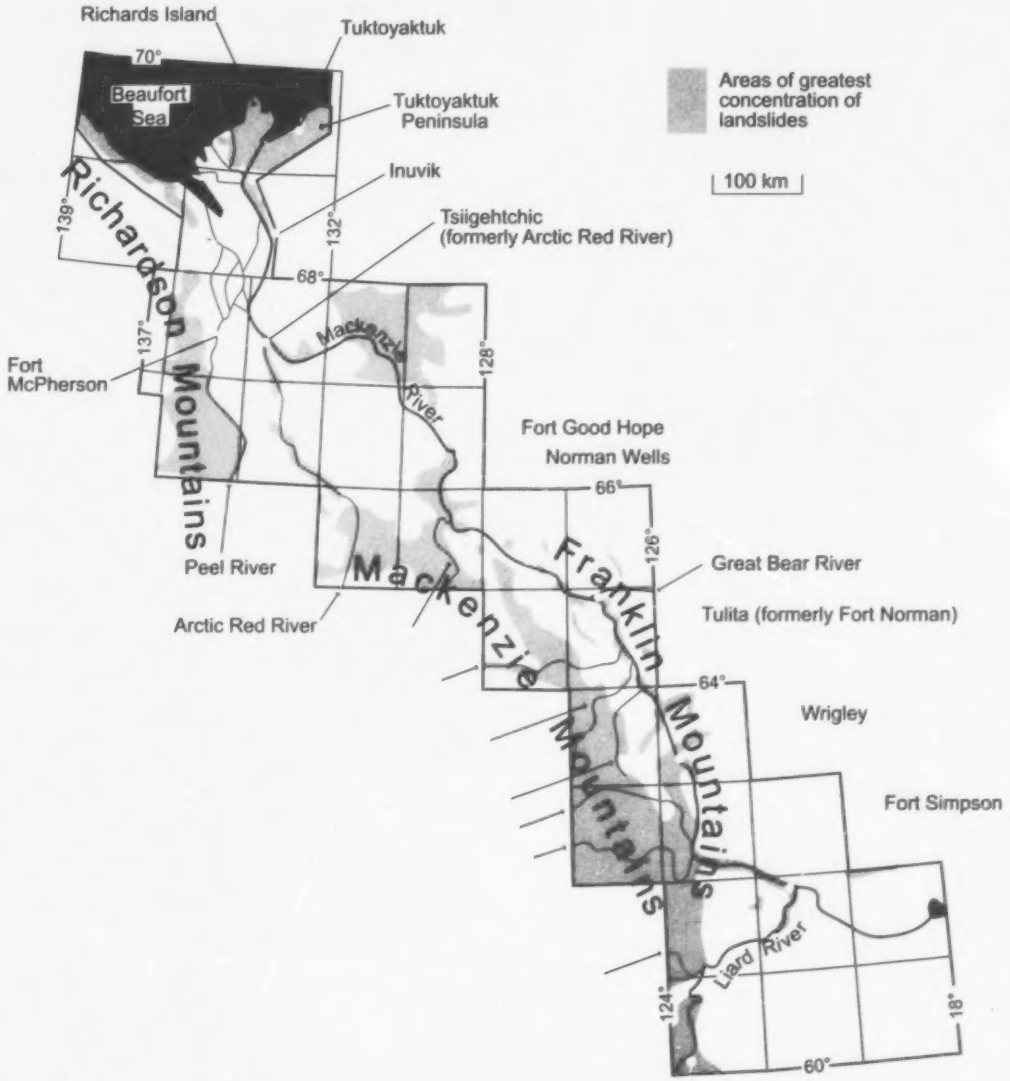

Figure 1. *Location map of study area depicting where landslides are most common.*

their mechanisms of failure vary depending on local relief, geology, permafrost and groundwater conditions. Many slope failures are closely related to the thawing of ice-rich permafrost or the pressure exerted by groundwater confined by permafrost. It is areas containing ice-rich unconsolidated sediments that would be most susceptible to landsliding resulting from the degradation of permafrost in the event of global warming.

In the Mackenzie valley and adjacent regions, the major impact of landslides will most likely be the disruption or destruction of pipelines, roads, and bridges. River navigation might be temporarily affected in the event of a landslide dam on a river. As well, landslide dams could cause further thermal instability upstream, possibly promoting additional landslides. Increased landsliding could also affect fisheries through possible increased siltation in streams and rivers, and possible destruction of spawning beds.

COMPILATION METHODOLOGY

Compilation of the landslide inventory is based primarily on the interpretation of airphotos or compilation from existing surficial geology maps (*see* listing of maps in Aylsworth et al., 2000). For the most part, the latter maps were also produced mainly from airphoto interpretation. North of 68° latitude, the interpretation is based on 1985 images, and 1948 to 1972 airphotos were used between 68° and 64°N. Between 64° and 60°N, most of the landslides in the area were compiled from existing surficial geology maps based on 1947 to 1971 airphotos; however, within a corridor extending approximately 3 km on either side of the Mackenzie and Liard rivers, all available airphoto images (1947 to 1990) were re-examined during this project. Throughout the entire study area, landslides that occurred subsequent to the years of airphoto coverage are only included in the inventory if reported either in field observations made by GSC staff, mainly during regional surficial geology mapping projects or permafrost studies, or from publications describing individual failures. Many of the slope failures along the banks of the Mackenzie River have been previously noted by Code (1973) and McRoberts and Morgenstern (1973), and, in the vicinity of the Beaufort Sea coast, landslides have been mapped and discussed by Mackay (1966, 1986) and Rampton (1988).

The inventory consists of 1:1 000 000 scale maps showing landslide distribution, differentiated by landslide type, and is accompanied by a digital database consisting of information on individual landslides or groups of landslides. Location, class and type of failure, dimensions, description of bedrock lithology or unconsolidated sediments, age if known, additional comments if any, and airphoto numbers are all given, and references if available. The database contains 2014 entries, of which 55% represent individual landslides and a further 10% represent a pair of closely spaced landslides. The remaining 35% of entries represent clusters of closely spaced landslides with numbers ranging from more than 2 to more than 30 landslides per cluster, with most clusters consisting of fewer than 10. For the most part, these clusters represent groups of small retrogressive thaw flows in the northern third of the study area.

CLASS AND TYPE OF LANDSLIDE

In the database, landslides were divided into those in which the failure plane occurred in bedrock and those in which the failure plane occurred in unconsolidated Quaternary sediments of glacial, glaciofluvial, lacustrine, marine, alluvial, or colluvial origin. In many cases it was the cementing presence of ice-bonded permafrost that gave strength to the unconsolidated sediments. Of the 2014 landslides or landslide clusters in the database, 31% occurred in bedrock and 69% occurred in unconsolidated sediments. The landslides have been categorized by 'class' (Fig. 2a, in pocket), which describes the basic style of failure: flow, slide, or complex, which incorporates elements of both (based on Varnes, 1978). Many of the landslides in the southern region, particularly those in the Mackenzie Mountains, remain unclassified. 'Class" is further subdivided into 'type' (Fig. 2b, in pocket), which refers to the specific kinds of landslide and implies not only style of failure, but also the specific physical attributes of the landslide. The names of failures associated with melting of frozen ground or ground ice, and therefore unique to permafrost regions, follow common permafrost terminology.

Landslides in permafrost terrain are characterized by two distinct classes: 'flows' and 'slides'. These classes are based on mechanism of failure, although the differences are also reflected in morphology. Flows have a fluid character, showing evidence of mobility throughout the failure. Slides, on the other hand, show evidence of rigid movement in that components of the slide move downslope as more or less intact blocks. Two other classes of failure mechanism, 'falls' and 'topples' (the names are self-evident), have a minor role, particularly as a contributing factor to the development of a larger failure, but are generally too small to distinguish on an airphoto and are not included in the inventory. 'Complex' is assigned when not only is there more than one type of movement, but also each type of movement accounts for a significant volume of the landslide. Two mechanisms of failure may be active at the same time in an individual landslide, or, as a landslide develops, one type of failure may evolve into another type. Slope movement in permafrost regions may continue either continually or cyclically for weeks, years, or even decades.

Several different types of landslides make up each of the two major classes, flows and slides. Flows can be subdivided into shallow, active-layer detachments (or skin flows), deeper retrogressive thaw flows (also commonly known as bimodal flows or ground-ice slumps), and rapid debris flows. Slides are divided into rotational slides and translational (or planar) slides.

Active-layer detachments

Shallow slope failures involving detachment and downslope movement of only the active layer and vegetation mat are known collectively as active-layer detachments (Fig. 3, 4, 5). Movement may involve either sliding of a relatively intact, thawed piece of ground on the underlying frozen sediment (sometimes known as an active-layer glide) or flow of water-saturated sediment (also known as a skin flow). They are generally triggered by unusually warm temperatures or some disturbance of the vegetation mat (*see* Dyke, 2000). Either instance will result in overdeepening of the annual active layer, which, if icy sediment is thawed, may detach and move downslope. The landslide tends to expand laterally as adjacent icy sediment also thaws. This type of failure is common

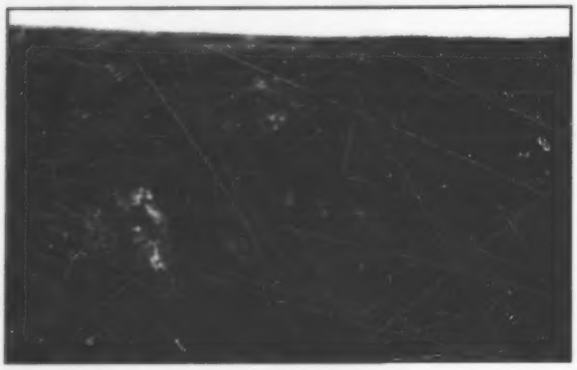

Figure 3. *Landslides along the Mackenzie riverbank near the mouth of Thunder River. These still-active flows, triggered following the 1986 fire, have evolved from shallow skin flows into flow complexes. The upper portion of the landslide has developed into a retrogressive thaw flow as continued retrogression in the newly exposed, ice-rich sediment has resulted in deepening and steepening of the headscarp. The rapid movement on the steeper part of the bank may more properly be called a debris flow. Photograph by J.M. Aylsworth in 1994. GSC 2000-034A*

Figure 4. *Head of active-layer detachment (skin flow) in the 1986 Thunder River fire scar. Photograph by J.M. Aylsworth in 1994. GSC 2000-034B*

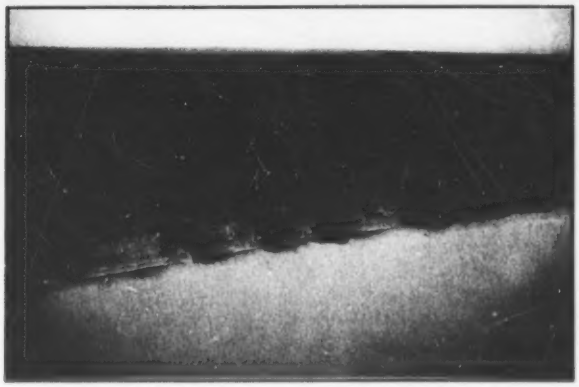

Figure 5. *Active-layer detachments (skin flows) line the banks of the Mackenzie River south of Tulita (formerly Fort Norman). Flows were triggered following a large forest fire in 1995. Photograph by M.M. Burgess in 1997. GSC 2000-034C*

in ice-rich terrain following a forest fire if the insulating organic mat is damaged or destroyed. Numerous active-layer detachments were reported in the fall and spring near the Mackenzie River following the summer 1994 and 1995 forest fires southeast of Tulita (formerly Fort Norman) (M.M. Burgess, K.L. MacInnis, pers. com., 1994, 1995). If the removal of the water-saturated sediment continually exposes massive ice or icy sediments at greater depth, an active-layer detachment may develop into a retrogressive thaw flow. Such is occurring in several of the above mentioned examples (M.M. Burgess, pers. com., 1995, 1996). On steep slopes, the active-layer detachment may develop into a debris flow.

Retrogressive thaw flows

The most common type of flow is the retrogressive thaw flow (Fig. 6, 7, 8). Retrogressive thaw flows occur only in ice-rich terrain. They have a characteristic bowl shape and a bimodal profile with a steep headwall and a low-angle tongue. The headwall gradually erodes upslope as massive ground ice or icy sediment thaws in the scarp, and the resulting water-saturated sediment flows downslope away from the scarp.

Retrogressive thaw flows may be initiated by any process or event that results in the exposure to melting of massive ice or icy sediment. For example, active-layer detachments may develop into a retrogressive thaw flow if ground ice is exposed. In the northern part of the study area, massive ice may become exposed in dilation cracks resulting from growth of ground ice, or in tension cracks formed as icy sediment deforms and creeps under its own weight. Exposure of the ice-rich scarp, which is essential to the development of thaw flows, may also be initiated at coastal sites or along scarps by block falls or topples.

Once the massive ice or icy sediment is exposed in the headwall, a seasonal cycle becomes established, beginning in summer when snow melts away from the scarp face. Water derived from snow melt and melting of ground ice helps to clean the scarp face of debris that may have accumulated the

Figure 6. *Retrogressive thaw flow demonstrating the classic bowl shape, with a steep, ice-rich headscarp and a very low-angle tongue, Tuktoyaktuk Peninsula. Photograph by J.M. Aylsworth in 1992. GSC 2000-034D*

Figure 7. *Two recent retrogressive thaw flows in an older retrogressive thaw-flow scar, Tuktoyaktuk Peninsula. Massive ground ice is visible in the headscarp. Photograph by J.M. Aylsworth in 1992.GSC 2000-034E*

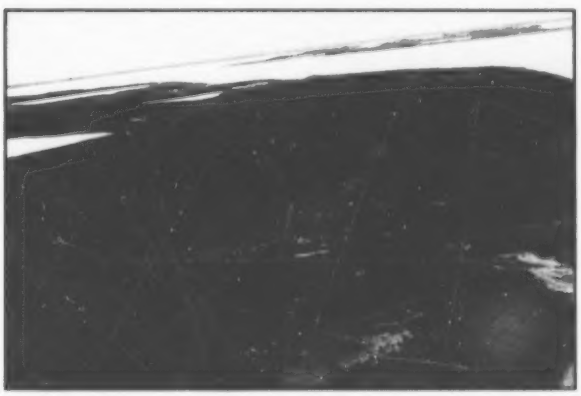

Figure 8. *Large retrogressive thaw flows, northern Richards Island. Extensive areas of the Beaufort Sea coast are subject to similar slope failure. Flows remain active as long as ice is exposed in the headscarp and debris is removed from the toe of the landslide by erosion. Photograph by J.M. Aylsworth in 1992. GSC 2000-034F*

previous season. As sediment is released from the melting ice, it falls, slides or flows down the scarp face. Melting of the ice undercuts the active layer at the top of the scarp, and the undercut peat and sediments detach and fall down to the base. These become incorporated into the water and debris at the base of the scarp face and flow downslope. Eventually the slide stabilizes as the face becomes buried by debris, the ice content of the sediment at the scarp decreases, or the slope of the scarp decreases (Mackay, 1966).

Retrogressive thaw flows tend to be cyclic in nature. The scarp face may become buried if debris is not removed at a rate faster than it is supplied. If the flow is not reactivated immediately, it can become quickly revegetated because any fallen blocks that remain intact carry vegetation into the bowl of the slide (Mackay, 1966). If the flow does reactivate, a new scarp face may develop within the old scar, and the old flow deposits are recycled in the new failure. In icy sediment, a retrogressive thaw flow will remain active wherever the slope is sufficiently steep to continually evacuate the scar, or where fluvial erosion transports debris away from the toe of the landslide, or in the case of a lake or coastal shoreline, if wave energy is sufficient to erode the flow debris and keep the scarp face exposed. Rising water levels and wave activity during a storm may reactivate a stagnated slide by eroding the toe of the slide and triggering more movement of debris away from the face (Carter et al., 1987). During major storm surges the debris may be overtopped or quickly eroded back to expose the ice face to direct wave attack.

Debris flows

Debris flows, rapid flows of water-saturated sediment, occur in areas of higher relief (Fig. 9, 10). These flows are typically long and narrow, and widen into debris fans at their base. They may be triggered by extreme precipitation or unusually deep seasonal thaw in an abnormally warm summer. For example, the abnormally warm summer temperatures of 1989

Figure 9. *Debris flow of 1990, Peel River. Flow, which occurred in till on the slopes in the distant background, extends for approximately 2 km to river level. Photograph by L.D. Dyke in 1992. GSC 2000-034G*

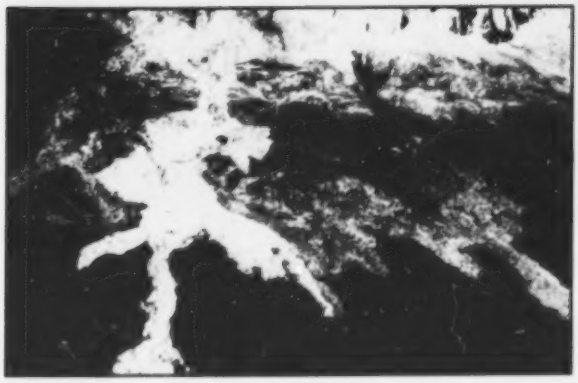

Figure 10. *Debris flow in colluvium at Nahanni Butte triggered by the 1988 extreme rainstorms in the Liard Basin Photograph by J.J. Clague in 1988. GSC 2000-034H*

Figure 11. *Shallow rotational slide in lacustrine silts failed as a single block sliding along a circular failure surface. West bank of the Mackenzie River south of Old Fort Point. Photograph by J.M. Aylsworth in 1994. GSC 2000-034I*

Figure 12. *Deep-seated rotational slide in frozen lacustrine sands and silts, Mountain River, occurred as a series of rotational block movements. Unfrozen clay occurs at river level. Photograph by J.M Aylsworth in 1992.GSC 2000-034J*

triggered a debris flow in the Caribou Hills east of the Mackenzie Delta that ran 1.5 km downslope in a few minutes (Aylsworth et al., 1992). Exceptionally heavy precipitation induced a debris flow at Nahanni Butte on the Liard River that narrowly missed a ranger station (Clague, 1988) and a large translational slide in bedrock along a tributary of Mountain River was reactivated after a heavy rain in the summer of 1971. In mountainous areas, active-layer detachments formed in burn scars may develop into debris flows if large amounts of meltwater are produced by the thawing of newly exposed ground ice.

Rotational slides

Rotational slides involve the downslope movement of a rigid block of sediment along a curving failure surface, such that the toe of the block extends well beyond the original slope and the original upper surface of the block is generally back-tilted towards the scarp of the landslide (Fig. 11, 12, 13, 14, 15, 16). Such failures may occur as an individual block or as a stepped series of blocks. Many multiple failures have occurred.

Rotational slides are commonly induced by undercutting of a riverbank by erosion. Pressure of ground water, confined beneath the permafrost layer, is probably an important contributor in these failures (*see* Dyke, 2000).

Rotational slumps may subsequently deform and move as flows, probably as ice within the slump melts. The flow may drastically modify the shape of the original feature. In this case the landslide would be classified as a slide-flow complex. Numerous examples of this type of landslide can be found along the banks of the Mackenzie River downstream of Fort Simpson and upstream of Tulita at Old Fort Point (Fig. 17; McRoberts and Morgenstern, 1973).

Figure 13. *Several rotational slides line the Mackenzie riverbank south of Old Fort Point in this airphoto. Arrows indicate block cracks behind the landslides. The fresh landslide occurred in July, 1984, as a series of block failures. Prior to 1984, the block cracks extended to the right along the present headscarp. Another large rotational slide occurred upstream (right) of the 1984 slide in the winter of 1997. NAPL A26839-125 in 1984*

Figure 14. *Large rotational slide occurred in February or March, 1997, immediately upstream of the 1984 landslide shown in Figure 13. Part of the 1984 landslide is visible in the extreme left of the photo. Photograph by M.M. Burgess in 1997. GSC 2000-034034K*

Figure 15. *Close-up of headscarp of the 1997 rotational slide. Landslide occurred in frozen lacustrine silt over clay. The shape of the feature has evolved since this photo was taken, as frozen blocks thawed. Photograph by M.M. Burgess in 1997. GSC 2000-034L*

Figure 16. *Large cracks, indicated by arrows, are visible on the flat surface behind the headscarp of the 1997 landslide. Photograph by M.M. Burgess in 1997. GSC 2000-034M*

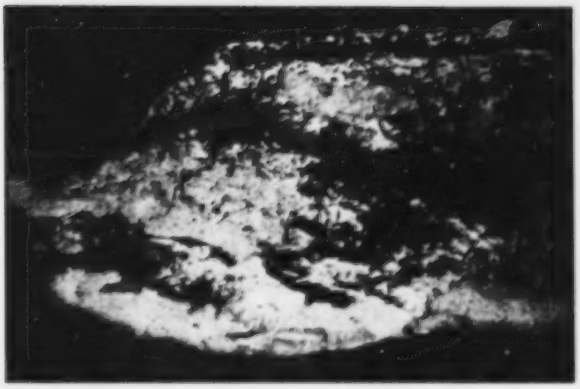

Figure 17. *This landslide, in lacustrine silts overlain by sands, occurred as a series of rotational blocks during the winter, 1992. Subsequent flow occurred as the sediment thawed. Photograph by J.M. Aylsworth in 1992. GSC 2000-034N*

Translational slides

Translational slides involve rigid movement along a planar failure surface, commonly a bedrock bedding plane. They may be induced by fluvial erosion of the base of an inclined bedding plane. These slides are restricted to mountainous areas.

LANDSLIDE DISTRIBUTION

Landslides occur the length of the Mackenzie valley; however, they tend to be concentrated in specific zones within the study area (Fig. 1, 2a, c, d, e). These zones include the banks of the Mackenzie and its tributaries where the rivers have eroded deep channels into the underlying unconsolidated Quaternary sediments, or, in places, bedrock; the slopes rising towards the Mackenzie and Richardson mountain fronts; the steep slopes of the ranges in the Mackenzie and Franklin mountains to the west of the Mackenzie River, and to a lesser extent the slopes of the ranges east of the Mackenzie River; the shores of the hundreds of small lakes of the Tuktoyaktuk Peninsula and Richards Island; and the cliffs of the Beaufort Sea coast.

Landslides range in area from 225 m^2 to 20 km^2. The mean area of a landslide in the inventory is approximately 0.5 km^2 and median value is 0.25 km^2. Area of a landslide is considered to include the area of the scar and of the debris tongue. In most cases it was roughly calculated from the mean width and length of the feature. Volume was not calculated. In general, size of landslide increases towards the south. The northern part of the study area is characterized by hundreds of small retrogressive thaw flows. Rotational slides, which generally are larger, are more frequent in the southern part of the study area. Many of the massive landslide areas are composed of several large coalescing and overlapping landslides which may represent years of landslide activity rather than one event.

Elevations of the scars range from a few metres above sea level along the Beaufort Sea coast to 1980 m in the Mackenzie Mountains. A general trend of increasing elevation of landslide scar to the west reflects the general rise to the Mackenzie and Richardson mountains. Although there is a range of elevations represented, generally landslides are restricted to lower elevations (500 m or lower). These are the areas with the greatest accumulation of unconsolidated Quaternary sediments, in particular glaciolacustrine sediments, and also the areas most affected by river erosion. Landslides at higher elevations are almost exclusively bedrock failures on steep slopes.

Although no strong relationship exists between landslide frequency and aspect (direction of movement), there are somewhat fewer landslides with a northerly aspect. When only landslides in unconsolidated Quaternary sediments are considered, a westerly aspect is slightly more dominant than the other directions. For the most part, the role of aspect unrelated to site control is insignificant. Site considerations, i.e. position with respect to erosion along a river or inclination of a bedrock formation, are more important.

Within the above-mentioned zones of dense landsliding, distinct types of slope failures are associated with specific geological units.

GEOLOGICAL ASSOCIATION

Certain geological materials are particularly prone to slope failure. In his survey of the Mackenzie riverbanks, Code (1973) noted that landslides involving bedrock occurred most commonly in weak, soft shale and weakly cemented Cretaceous sandstone and siltstone, where any steep bank exceeding 30 m height was potentially unstable. He noticed for fewer landslides in the mainly well cemented, resistant Devonian rocks along the river. However, further west, in the ranges of the Mackenzie and Franklin mountains, numerous landslides have occurred in rock that is Devonian or older.

Regardless of geological age, most landslides in rock occur in shale or in shale-mudstone-siltstone formations — soft, thinly bedded and generally friable and fissile sedimentary rock. For example, numerous landslides scar slopes in the shales of the Fort Simpson Formation, which is widespread in the southwestern part of the study area, as well as in other formations such as the Clausen and Mattsen. Where landslides occur in harder, more competent rock such as limestone, dolomite, sandstone, and quartzite, sliding usually occurs along the bedding plane. For example, large landslides are found in the limestones and dolomites of the Nahanni, Funeral, and Whittaker formations, and the Arnica and Sombre formations, respectively. Yet in most cases involving more competent bedrock, the failure plane is actually in a shale bed within the formation such as the Trout River and Redknife formations (sandstone and siltstone with shale beds) or the Flett formation (limestone with shale), or the failure plane lies in an underlying shale formation. Shale may move as a flow (active-layer detachment or debris flow) if very moist and disintegrated or as a slide; limestone and dolomite fail as a slide, either a translational failure along a bedding plane, or a rotational failure on underlying weak shale. Bentonite layers have been identified as crucial to some failures (K.W. Savigny, pers. com.,1992).

For unconsolidated Quaternary deposits (Fig. 2e), the greatest number of landslides (65%) occurred within till, a glacial sediment having a wide range of particle sizes. This number of failures is probably a reflection of both the potentially icy nature of the fine-grained matrix and the widespread distribution of till as compared to other Quaternary sediments within the region. In some cases, the failure plane is at the till/bedrock interface. Lacustrine clays and silts are involved in 17% of the landslides, and glaciofluvial sands, silts, and gravels (10% of the landslides), deltaic sands and silts (5% of the landslides) and other deposits (3% of the landslides). In the case of glaciofluvial and deltaic sediments, the deposit commonly overlies lacustrine fine-grained sediments and in many instances the failure includes these underlying sediments.

Fine-grained sediments (silts and clays or fine-matrix tills) are commonly ice rich, with ice content varying from ice crystals to massive lenses of pure ice. These silts and clays are found in parts of the Mackenzie and adjacent valleys that were temporarily flooded by glacial lakes during retreat of the Laurentide Ice Sheet (*see* Duk-Rodkin and Lemmen, 2000; Aylsworth et al., 2000). High ice contents may also occur in fine-matrix till and fine-grained colluvium. Thaw in ice-rich, fine-grained sediments causes flow-type landslides, due to the low strength of the resulting water-saturated sediments. In these icy sediments, even low-angle slopes in areas of relatively little relief are susceptible to active-layer detachments and retrogressive thaw flows. If relief is sufficient, these flows can move very rapidly as debris flows. Although much less common, sands, if well saturated, also may flow rapidly downslope as debris flows.

Coarse-grained sediments (sands and gravels) are less likely to contain ice in excess of the pore volume. Landsliding is most likely along steep riverbanks eroded in sands and gravels deposited along the margins of the above-mentioned glacial lakes. These sands and gravels tend to fail in blocks as rotational slides. These failures also involve the underlying lacustrine sediments, and, in fact, are promoted by the presence of weak, unfrozen clays beneath the typically frozen sands. In many locations south of about 65°N latitude, the height of the riverbank and the thickness of the coarse-grained sediments is greater than the depth of permafrost, so that groundwater is free to move toward the bank through the sand and gravel between the overlying permafrost and underlying clay layer. If the cliff face is stable, a frozen seal of ice-cemented sediments may develop on the slope and artesian pressures build up, triggering a failure. Conditions such as rapid erosion by the river of the base of the bank or insulation by deep snow cover at the base of the bank can remove that permafrost plug. The resulting rapid drawdown of water in saturated sediments may induce a landslide by initiating erosion of loose sediment.

IMPLICATIONS OF CLIMATE CHANGE

In permafrost regions, climate warming is likely to reduce the stability of slopes by causing a reduction in the thickness and the strength of frozen ground. However, it is presently not possible to predict the precise impact of climate warming because of the lack of data on the failure strength of frozen and thawing sediments and bedrock. In Dyke (2000) the sensitivity of slope stability to decreasing permafrost thickness or increasing active-layer thickness is discussed. This analysis suggests that thawing of icy permafrost sediments would significantly reduce slope stability.

Rockslides are less likely to be affected by climate change. They are more commonly induced by random physical processes such as earthquakes or river erosion, or by weathering and loss of strength along existing planes of weakness. Many may have occurred following glacial retreat as a result of oversteepening of valley walls by glacial erosion.

Although climate warming will reduce slope stability, it will most likely act only to accelerate an already active process. Many rotational slides are triggered by removal of the slope toe and undercutting of the bank by river erosion, which not only causes failures, but can maintain movement of an active slope by continually removing the supporting debris that accumulates at the base of the slope. Wave action in lakes and along coasts plays a similar role. On the other hand, anything that deepens the active layer, whether it be abnormally warm air temperature, removal of the insulating organic mat by fire or other means, or thermal instability resulting from proximity to warm water bodies, can cause slope failure. In icy sediment, the first two conditions would result in an active-layer detachment which may then develop into a larger failure (retrogressive thaw flow). The third condition, which would cause loss of support at the base of a slope, may induce a larger, deep-seated failure. Also, pore-water pressure from groundwater confined by permafrost may become more effective as permafrost thins (*see* Dyke, 2000).

If climate change should occur in the Mackenzie valley as suggested, then there should be warmer air temperatures, particularly in winter and spring, drier conditions due in part to increased temperatures, decreased snow and an earlier melt, and more extreme storms (consensus of opinion at the Mackenzie Basin Impact Study Final Workshop, Yellowknife, 1996). The ground temperatures would respond to these changing climate conditions; permafrost would degrade, possibly disappearing altogether in the south and active layers would deepen, melting interstitial ice, elsewhere. Given the prevalence of permafrost and icy sediments in this area, any regional ground-temperature warming that might occur would increase the frequency of landslides in the area, at least in the short term. Active-layer detachments and retrogressive thaw-flow slides are the most likely failure types under those conditions. Likewise, any increase in extreme storms will result in an increase in landslides. Heavy rainfall may saturate the ground, favouring debris flows, or increase the pore-water pressure in slopes, leading to deep-seated rotational slides. Riverbanks may be undercut by higher discharges in rivers following these storms, inducing rotational slides or smaller topples. The latter, by exposing icy sediment, may develop into retrogressive thaw-flow slides. Drier conditions, making forests more susceptible to lightning strikes, will result in fire-induced, active-layer detachments. In the north, rising sea level will inundate coastal and estuarine areas, creating thermal instability at the base of frozen slopes.

At a regional scale, those areas most susceptible to landslides in the event of climatic warming are likely to be areas that presently are experiencing significant numbers of landslides. At a more localized scale, terrain sensitivity will be dependent on local geology, topography, permafrost, and ground-ice conditions. Sites most susceptible to climatically induced landslides include ice-rich, fine-grained sediments on slopes in close proximity to bodies of water or rivers, or frozen coarse-grained sediments overlying clay or clayey till in steep sections along riverbanks where permafrost does not extend to the base of the section. In addition, fire scars on ice-rich slopes are susceptible to landsliding.

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Stability of permafrost slopes in the Mackenzie valley

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Abstract: Slope stability is conventionally analyzed in terms of a 'factor of safety', that is, the ratio between the shearing strength of a material forming a slope and the forces tending to cause sliding or failure of that material. In permafrost regions, many failures are triggered by a decrease in strength of frozen ground due to excessive thawing and deepening of the active layer. This results in an active-layer detachment. This type of slide is often triggered by excessive thaw following the destruction of surface vegetation by fire. The most likely impact of climate warming may be an increase in active-layer detachments if there are more fires. Although most of the deeper rotational failures occur along river banks where erosion at the foot of the slope is an important factor, long-term decreases in permafrost thickness could also contribute to instability.

Résumé : Par convention, la stabilité des versants est analysée en fonction d'un « facteur de sécurité », c'est-à-dire du rapport entre la résistance au cisaillement des matériaux qui constituent un versant et les forces qui ont tendance à causer des glissements ou des ruptures dans ces matériaux. Dans les régions pergélisolées, de nombreuses ruptures sont déclenchées par un abaissement de la résistance du gélisol à cause d'un dégel excessif et une augmentation de la profondeur du mollisol. Il en résulte un décollement du mollisol. Ce type de glissement est souvent provoqué par un dégel excessif attribuable à la destruction de la végétation par un incendie. La répercussion la plus probable d'un réchauffement climatique pourrait être une augmentation des décollements de mollisol associés à une augmentation des incendies. Même si la plupart des ruptures rotationnelles profondes se produisent le long des berges des cours d'eau, là où l'érosion au pied des versants est un facteur important, les diminutions à long terme de l'épaisseur du pergélisol pourraient concourir davantage à l'instabilité des versants.

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INTRODUCTION

Approximately 2000 landslides in the Mackenzie valley between Inuvik and Fort Simpson and about 1000 in the ground-ice rich coastal uplands near the Beaufort Sea have been mapped from airphotos (*see* Aylsworth et al., 2000). This abundance of slides in a permafrost region with abundant ground ice suggests that thawing of this ice plays a role in landslide occurrence. Therefore, climate warming becomes an additional factor contributing to slope failure in this region because of the potential it holds for melting ground ice. This paper examines how permafrost slopes in the Mackenzie valley fail and analyzes the sensitivity of slopes to factors controlling slope stability.

Whereas another paper in this bulletin (Aylsworth et al., 2000) on landslide distribution gives a comprehensive classification of landslides, this paper concentrates on two mechanisms which often initiate slope failures in or on permafrost in the Mackenzie valley: deep seated rotational failures and active-layer detachments (Aylsworth et al., 1992). Rotational slides involve failure through permafrost and may be triggered by a decrease in permafrost thickness. Active-layer detachments form along the top of permafrost and are triggered by excessive amounts of water produced by rapid or particularly deep active-layer development in ice-rich ground. Both types of failure often evolve into more complex features as thawing of the slide mass or adjacent exposed ice-rich sediment continues.

DEFINITION OF SLOPE STABILITY AND FACTOR OF SAFETY

Landslides occur because the forces resulting from the weight of the material forming the slope overcome the shear strength of that material. The conventional engineering approach to assessing the stability of a slope is to compare these two opposing factors. The proportion by which the strength exceeds the forces tending to overcome this strength gives a measure of the stability and is termed the 'factor of safety' against failure (FS):

$$FS = \frac{\text{Strength of slope material}}{\text{Shear stress acting along failure surface}}$$

If FS is greater than 1, the slope is stable; if equal to or less than 1 it is quasi-stable or unstable and should fail.

Determining the forces due to the weight of material is the easiest part of a stability prediction because soil or rock unit weights can be accurately measured and variability in unit weight can be assessed from surface samples or compiled from boreholes. Accurately determining soil strength and the location of the surface along which failure will take place is much more difficult. Laboratory techniques for measuring material strength are routine, but the accuracy with which the results represent the strength mobilized in the ground is difficult to know. Furthermore, the position and shape of the failure surface along which this strength is mobilized must be estimated. Without other constraining information it is usually chosen as the surface calculated to give the lowest factor of safety. Finally, the presence of water in the slope material is fundamental to determining stability. Pressure in soil pore water can be high enough to counteract the weight of overlying material and eliminate the friction which may be maintaining the slope. This factor is particularly important in ice-rich ground where thaw can produce considerable quantities of pore water. Uncertainties in the magnitude or nature of these variables lead to the selection of safety factors considerably in excess of 1 when slopes must be designed as part of an engineering project.

LANDSLIDE MECHANISMS

'Rotational slides' involving deep-seated failure in rock or Quaternary sediments form along concave or dish-shaped failure surfaces. The slide mass rotates as a rigid body, leaving a headscarp steeper than the original slope. The lower part of the slide mass usually moves out onto the base of the slope. The factor of safety against this kind of failure is customarily determined by representing a slope by a single vertical cross-section. This approach simplifies the problem by implying that the cross-section is representative of the wider section of slope which will almost certainly be involved in a failure. The simplest analysis assumes a circular failure surface. The slide mass rotates as a rigid body without undergoing any internal deformation, requiring only the shearing along the failure surface to be considered in the factor of safety analysis. Also, the analysis only predicts initial slope failure; it does not provide any indication of how far the slide mass will move down the slope.

Failure usually begins as a rigid body movement with internal deformation of the slide mass occurring once the toe moves out beyond the base of the initial failure surface (Fig. 1). In an extreme case, slide masses may become completely fluidized due to high moisture contents and melting ground ice and move easily on angles much less than that of the initial failed slope. This results in a complex failure in which the fluid behaviour can modify considerably the initial rigid form. While prediction of fluid movements is more difficult, some indication of subsequent movement will be given by sediment consistencies, ice contents, and length and steepness of slopes beneath sites of initial failure.

The other failure mode discussed is the 'active-layer detachment'. This involves only the active layer with the failure surface determined by the boundary between thawed and frozen sediment. Stability can be analyzed by the infinite slope method. With this technique, the failure surface is considered as a planar surface that is long compared to the slide thickness. The added resistance where the failure surface cuts across this thickness can thus be neglected. The mat of living vegetation and roots that usually exists at the ground surface can be the strongest part of the slide mass and is often found compressed and crumpled at the base of a slide. This kind of slide may deepen and lengthen if thaw continues into ice-rich ground exposed by the original slide (Fig. 1). If this happens, it becomes a retrogressive thaw flow.

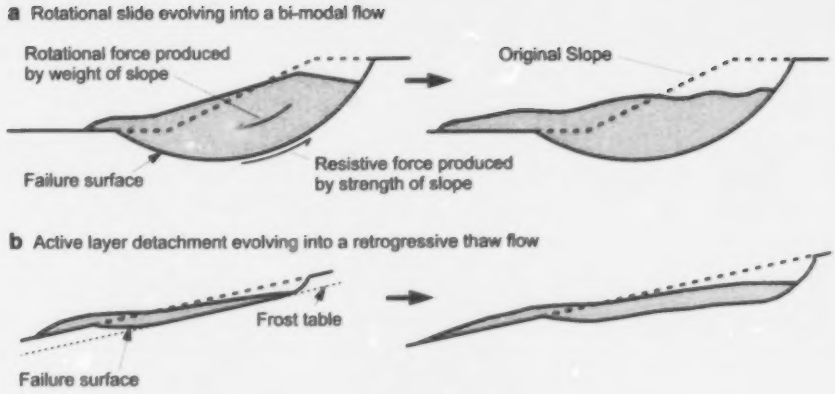
a Rotational slide evolving into a bi-modal flow
Rotational force produced by weight of slope
Original Slope
Failure surface
Resistive force produced by strength of slope
b Active layer detachment evolving into a retrogressive thaw flow
Frost table
Failure surface

Figure 1. *Two primary types of landslides occurring in permafrost in the Mackenzie valley.* ***a****) Deep-seated rotational slides take place where a frozen layer cannot be supported by an underlying weaker material. This kind of slide may evolve into a flow if the slide mass is ice-rich.* ***b****) Active-layer detachments take place where an excessively deep thaw encounters ice-rich ground. Deep thaw may be caused by a particularly warm summer or the destruction of insulating surface vegetation. If thawing continues to encounter ice, a retrogressive thaw flow may develop.*

TRIGGERING OF SLIDES

Factors of safety are only as accurate as the strength and weight measurements used to calculate them. Because of this uncertainty, a slope that is stable according to the calculated factor of safety can still slide. Therefore, it may be more useful to recognize situations where the factor of safety will be reduced. If an ice-bonded layer thaws or a slope is undermined by river scouring, strengths will decrease or forces will increase. While the factor of safety may be uncertain, these situations are definite indications that stability is deteriorating. Eventually, the factor of safety may drop below 1. There are a number of processes which may trigger slides and the sensitivity of slopes to strength or loading changes caused by these processes will be discussed with respect to specific features.

ROTATIONAL SLIDES

Large rotational landslides are common along the Mackenzie River and some major tributaries. They occur in banks where these streams have cut through glaciolacustrine sediments or into poorly indurated rocks, such as the Cretaceous shale units along the Peel River and its tributaries. Banks can be 60 m or more high. Undercutting by scouring and toe erosion, especially on bends, maintains slopes close to the angle of repose where cohesionless sands and gravels are encountered. Although scouring may be a triggering mechanism, especially if undermining occurs, other processes can contribute to reducing the factor of safety. In the case of rotational slides, a decrease in the thickness of ice-bonded sediment or an increase in pore-water pressures beneath ice-bonded sediments will accomplish this. Therefore steepening of the slope angle by river scouring may, by itself, be insufficient to initiate a slide. Other processes occurring in conjunction with toe erosion may actually trigger a slide.

The response of the factor of safety to changing slope conditions can be studied by varying those conditions and noting how the factor of safety responds; however, to perform this type of analysis, the strength of the sediments forming the slope must be known. An added complication is the scarcity of strength determinations on frozen sediments. An example of how this difficulty can be overcome is given by analyzing a recent rotational slide near the mouth of Mountain River (Fig. 2), a location where slope stability investigations have previously been undertaken (McRoberts and Morgenstern, 1973, 1974a).

Immediately upstream from the mouth of Mountain River, the outside of a broad bend impinges against a slope cut into glaciolacustrine and glaciofluvial sediments deposited in glacial Lake Mackenzie (Smith, 1992). Almost the entire bank for several kilometres upstream exhibits slope failure (Fig. 3). Most slides appear on the earliest air photography of the area (1950) but the recency of forest establishment on scars suggests that slides do not predate about 1900; however, two failures have occurred in the early 1990s. One has resulted from piping with subsequent subsidence of overlying frozen sediment (Burton et al., 1995). The other is a multiple rotational failure involving frozen sands overlying clays (Fig. 4).

The recent rotational failure occurs in a 50 m high bank and consists of three retrogressive blocks. The toe is outlined for about 300 m along the edge of Mountain River by a series of low upthrust ridges of clay. These upthrusts, together with

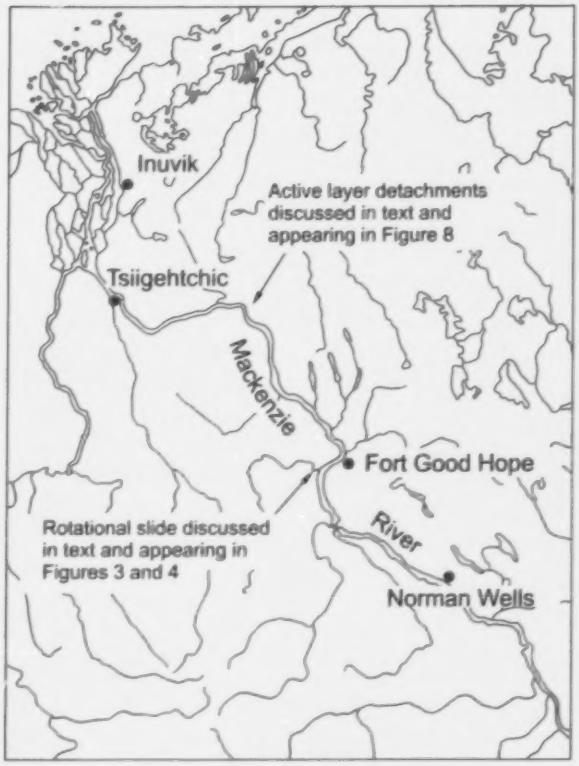
Inuvik
Active layer detachments discussed in text and appearing in Figure 8
Tsiigehtchic
Mackenzie
Fort Good Hope
Rotational slide discussed in text and appearing in Figures 3 and 4
River
Norman Wells

Figure 2. *Location of landslides discussed in detail.*

the obvious downdropping of the top of the slope and rotation of the blocks, indicate that the failure has occurred along a curved surface extending from the main headscarp to a considerable depth below river level, then returning to the surface at the upthrust ridges. The position of the toe and the headscarp, together with the headscarp angle enables the

Figure 4. *Rotational failure near the mouth of Mountain River. Slump blocks, consisting of deltaic sands deposited in glacial Lake Mackenzie by the ancestral Mountain River, are indicated by the patches of tilted forest. The headscarp, rotation of blocks, and river-edge uplifting of sediment all imply a circular failure surface that extends well into unfrozen clays below river level. Photograph by L.D. Dyke. GSC 2000-035A*

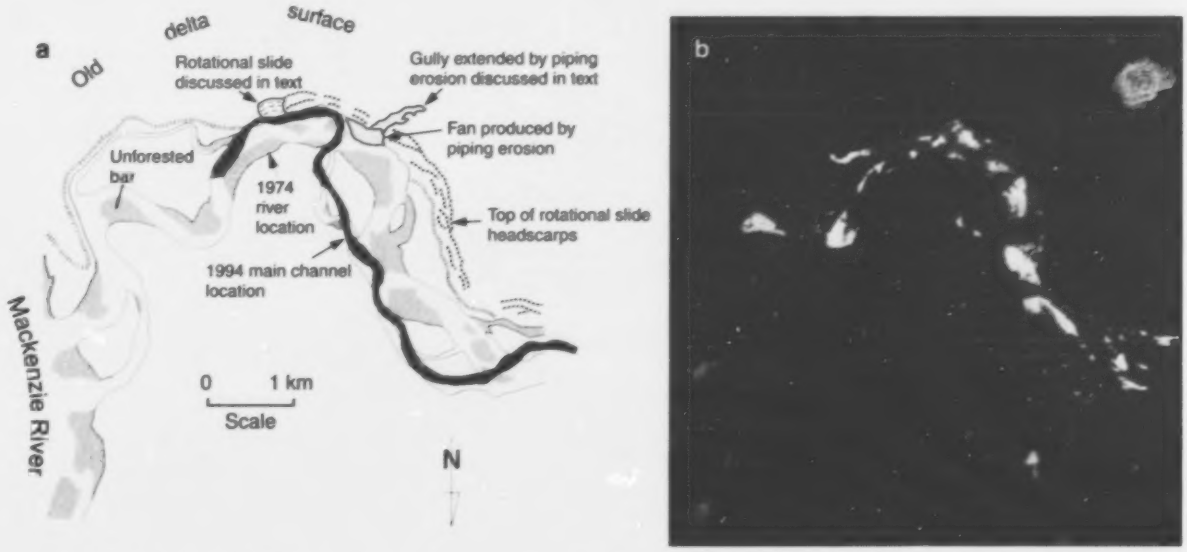

Figure 3. ***a)*** *The lowest reaches of Mountain River as traced from the adjoining* ***b)*** *1974 airphoto (NAPL A21528-223) and the main channel position in 1994. Continual scouring over this time interval has probably triggered the rotational slide discussed in the text. Hence, stability in part depends on channel changes in the river. This situation is typical where tributaries and the Mackenzie River are incised into glaciolacustrine sediments. Deep-seated rotational slides are favoured because of the weakness of unfrozen clays which typically underlie frozen sands.*

failure surface to be located if it is assumed to be an arc of a circle (Fig. 5). With a failure surface specified, strength at failure can be estimated.

Failure strength of slope materials

The failure strength of geological materials consists of a frictional and a cohesive component. The 'frictional component' is a measure of how abruptly an increase in weight on the potential failure surface increases the resistance to movement. It can be likened to the angle (ϕ) required to initiate sliding of a block down a slope. The weaker the frictional component, the lower the angle required for movement. If the pressure of water occupying the pores (pore-water pressure) of a soil rises towards the pressure from the weight of overlying soil, the frictional strength will tend toward zero ($\phi = 0$), reducing the factor of safety and making failure possible on increasingly shallow angles. The 'cohesive component' is the strength imparted by bonding or cementation between mineral grains of the soil. Cohesion is usually low for coarse-grained glacial sediments, i.e. sands and gravels.

At the Mountain River site, glaciofluvial sediments, consisting primarily of sands, overlie glaciolacustrine clay. The sands are frozen to a depth of 50 m (McRoberts and Morgenstern, 1974a), leaving unfrozen sand and clay below this depth. Clay is exposed only at river level but is thought to extend to the depth reached by the failure surface. This stratigraphic sequence is consistent with the regional stratigraphic character of glacial Lake Mackenzie sediments (*see* Smith, 1992). The low strength of the unfrozen clay under the frozen sand permits the slope to fail. If both the sand and the clay were frozen, failure at this site would be very unlikely.

If the strength of the sand and the clay are known, then the sensitivity of the slope to changes in permafrost thickness can be determined. The values shown in Table 1 below have been assumed, measured, or calculated.

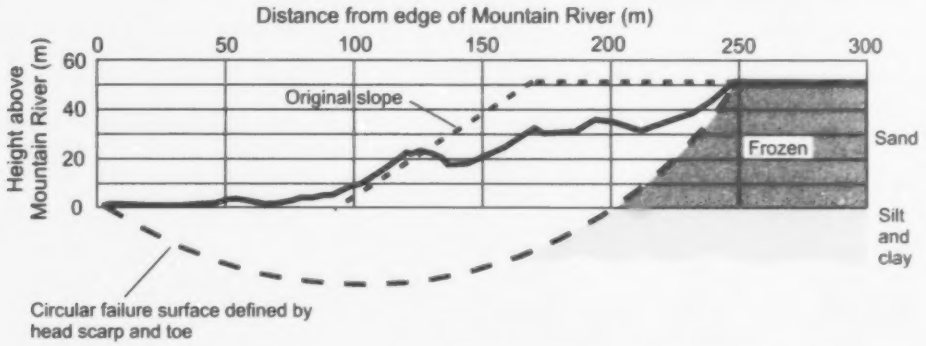

Figure 5. *A topographic profile of the Mountain River slide, surveyed in 1992, and the probable location of the main failure surface. The failure surface is defined by the location of the headscarp, the location of uplifted clay at the river edge, and the angle of the failure surface at the headscarp as implied by the location of slump blocks.*

Table 1. Values for strength components used in the stability calculation for the Mountain River site.

Material	Strength component	Description
Unfrozen clay	Frictional component	It is assumed that $\phi = 0$. Although ϕ is probably higher than this, a low value is likely, otherwise failure would not be possible along such a deep surface.
	Cohesive component	Measured at the Mountain River Slide on exposed unfrozen clay: 600 lb/ft^2 (0.29 kg/cm^2).
Unfrozen sand	Frictional component	It is assumed that $\phi = 30°$, a typical value for sand.
	Cohesive component	Assumed to be 0 because sands usually have little or no cohesion.
Frozen sand	Frictional component	It is assumed that $\phi = 30°$ because the friction should not be appreciably affected by the ice bonding and 30° is a typical value for unfrozen sand.
	Cohesive component	Calculated by finding the value that will maintain the slope at FS = 1 for the failure surface shown in Figure 5: 11 000 lb/ft^2 (5.4 kg/cm^2).

The cohesion of the frozen sand is found by conducting a stability analysis for the slope and stratigraphy shown in Figure 5. It is the only component of strength for which a value has not been assumed or measured. The cohesion of the frozen sand is determined by submitting trial values for the stability analysis until a value of FS = 1 is reached. Its value, as determined by this method, is listed in Table 1.

Sensitivity of slope stability to changes in slope conditions

The results of the strength analysis demonstrate that the frozen sands make by far the greatest contribution to stability at the Mountain River site. If bedrock occurred at river level, failure would be very unlikely. A factor of safety greater than 3 is indicated in Figure 6 for this case (i.e. depth of clay below river level = 0); however, the factor of safety rapidly decreases as the thickness of clay below river level is increased. This response points out the effectiveness of an underlying material with low strength in reducing the stability of an otherwise competent slope.

The importance of ice-bonding in maintaining stability suggests that stability is sensitive to thawing of ice-bonded permafrost. Climate warming might be expected to melt the top of an ice-bonded layer but groundwater flow beneath is also capable of transferring heat for thaw. Groundwater flow could dominate the thawing because this agent of heat transfer, in contrast to an air heat source, may act continuously in contact with the thawing surface; however, an analysis of this process would require knowledge of the degree to which climate warming would be reflected in groundwater temperatures.

The decrease in stability as permafrost thaws is shown in Figure 7. The figure is constructed on the basis that, before failure, the slope was barely stable (FS = 1) with permafrost at its present thickness. Without specifying whether thawing is taking place from above or below, it shows how the factor of safety decreases as permafrost in the sand becomes thinner. Plotting values of FS less than 1 shows that stability is quite sensitive to this change, suggesting that a few metres of thawing reduces the factor of safety appreciably. This analysis further suggests that slopes of this steepness and height were

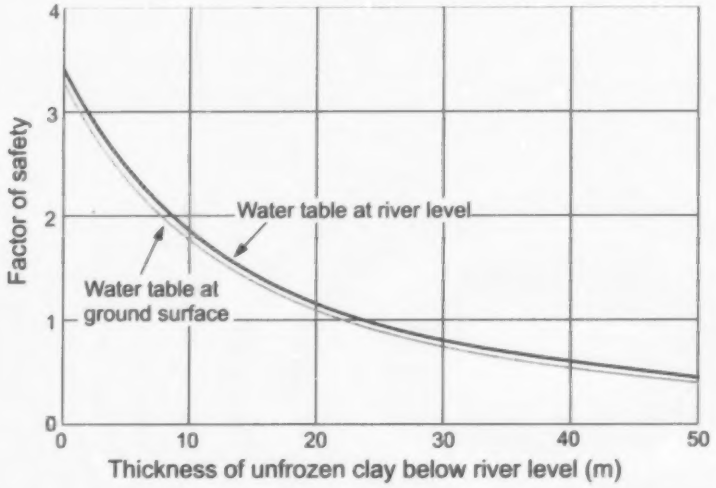

Figure 6.

The influence of unfrozen clay thickness below river level on slope stability at the Mountain River slide. Failure is very unlikely if competent bedrock lies immediately beneath the frozen sand; however, an increasing thickness of unfrozen clay reduces stability because a firm base is being replaced with a material having only a low frictional strength. Because the assumption of $\phi = 0$ implies a high pore-water pressure, it is almost equivalent to having the water table coincide with the ground surface. Therefore, the reduction in stability due to a high water table is small.

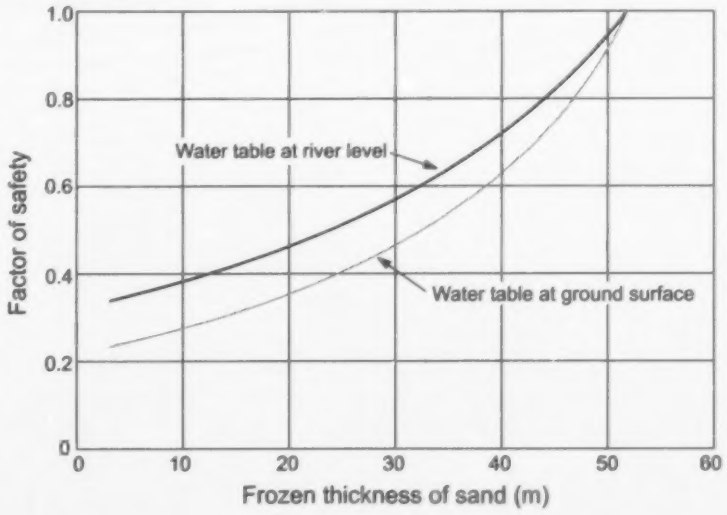

Figure 7.

The influence of permafrost thickness on slope stability at the Mountain River slide. Decreases in thickness of 5–10 m, probable in the event of 4–5°C of warming in the mean annual air temperature over 50 years (see Burgess et al., 2000), would reduce stability. Groundwater pressure equivalent to the pressure with the water table at the ground surface would further reduce stability if thawing was mostly from the bottom up.

entirely unstable before ice-bonding developed to the present depths. The present slopes may have been able to develop only after permafrost was established.

ACTIVE-LAYER DETACHMENTS

Failures allowing movement of the sheet of thawed sediment in the active layer are called active-layer detachments. Because they are difficult to detect on airphotos or they quickly evolve into retrogressive thaw flows, few are noted on the landslide distribution map in Aylsworth et al. (2000); however, they are probably one of the commonest types of landslide in the Mackenzie valley. Glacial and glaciolacustrine sediments are most susceptible to active-layer detachments, but weathered shale also exhibits this kind of failure. Once such a slide has exposed mineral soil, thaw will accelerate and reach greater depths than previously. If thaw continues into ice-rich ground, the original active-layer detachment may evolve into a retrogressive thaw flow where the slide deepens and develops a progressively thawing and retreating headscarp. This type of slide may not stabilize until the slide crater fills up with debris, the retreating headscarp reaches flat ground, or the scarp reaches ice-poor sediments.

The most noteworthy feature of active-layer detachments is the low slope angle on which sliding can take place. Many failures occur on slopes of less than 15° and failures on slopes as low as 3° have been recorded. These low failure angles remained unexplained by conventional stability analysis techniques until the role of excess pore-water pressure from thawing ground ice was recognized (McRoberts and Morgenstern, 1974b). If, during the summer thaw, water from melting ground ice is produced more quickly than it can escape to the surface, then the pore water begins to support the weight of overlying thawed sediment and the pressure in this water rises. Ice in excess of that needed to just fill the sediment pores and a rapid thawing rate can result in the pore water carrying the entire soil weight and hence a total loss of friction. If there is little or no cohesive strength, then high pore-water pressures available from thawing excess ice can induce slides on very shallow angles.

Active-layer detachments are most likely to be triggered by an event or condition that permits excessive pore-water pressure to develop in the active layer. During freeze-back, the active layer generally becomes ice-rich near the top as moisture from deeper levels is drawn to the advancing freezing plane. During thaw, moisture may move downwards to refreeze in the lower part of the active layer or migrate into the top of permafrost, producing an ice-enriched zone at the base of the active layer or in the top of permafrost (Mackay, 1980). An excessively warm summer may cause a higher than normal rate of melting. In this situation, failure is more likely as thawing reaches the lower levels of the normal active layer or extends into the ice-rich permafrost. This is because of the increasing weight of thawed material available to overcome any cohesion of the ground.

The event most likely to trigger an active-layer detachment is a forest or tundra fire. The heat from the fire, although intense, is short lived and produces little or no warming at the frost table (Mackay, 1977). Destruction of ground-surface vegetation, underlying litter, and peat is by far the most important effect because of the loss of insulation that this causes. The active layer subsequently deepens, eventually reaching a new equilibrium. This may take several years if the sediment is ice-rich to a considerable depth. If an active-layer detachment occurs, the bare soil surface will result in even greater thaw. Eventually revegetation will halt and ultimately reverse this increase; however, the time taken for revegetation will depend on the extent of further instability (i.e. retrogressive thaw flows) that develops after the initial active-layer detachment.

Sensitivity to failure

The tendency for an active-layer detachment to form can be determined by the same approach that was used for the rotational failure. In the case of the active-layer detachment, the failure surface is idealized as an inclined plane of indefinite (or infinite) extent. The fact that the failure plane must start at the surface is often ignored because this simplifies the calculation of the factor of safety. There is a force required to rupture the surface vegetation mat but it is probably very small compared to the resistance to movement along the rest of the failure surface. As with the rotational failure, the infinite slope analysis requires that the cohesion and frictional strength of the sediment be known (Hanna and McRoberts, 1988). Frictional strength can be greatly reduced while ground ice is melting and this phenomenon has been taken into account by McRoberts and Morgenstern (1974b). They incorporate a term called the "thaw consolidation ratio" into their analysis. This ratio compares the rate of melting with the rate at which excess water pressure is dissipated. The higher this number, the more the frictional strength is reduced.

Most of the Mackenzie valley has probably been burned at least once in the last 200 years. Many areas burned within the last 30 years are visible from the Mackenzie River. To illustrate the sensitivity of permafrost slopes to active-layer detachments and discuss the affect of forest fire on the thermal state of the ground, an example of this type of failure from the Thunder River area (approximately midway along Mackenzie River between Tsiigehtchic (formerly Arctic Red River) and Fort Good Hope, Fig. 2) is analyzed. Thirteen hundred square kilometres of forest adjacent the Mackenzie River in this area were burned in 1986 and within one year had developed extensive active-layer detachments (Harry and MacInnes, 1988). Several of these have evolved into retrogressive thaw flows (Fig. 8, 9). Active-layer detachments also developed within weeks of the extensive fires in the Tulita (formerly Fort Norman) area in 1994 (Savigny et al., 1995).

An example of the consequence of burning the vegetation mat is shown in Figure 10. This example shows a one year record of temperature at the base of the organic layers in adjacent burned and unburned sites in the Thunder River area. For

Figure 8. *Active-layer detachment slide in the Thunder River area. Bare mineral soil has been exposed by the removal of a thawed layer approximately 1 m thick. Photograph by L.D. Dyke. GSC 2000-035B*

Figure 9. *Retrogressive thaw flow slide in the Thunder River area. Ice-rich sediment can be seen in the head scarp of the slide. Thawing of the ice causes the slide to deepen and extend uphill, at the same time providing water to fluidize the sediment. Continued thawing of ice-rich mineral soil exposed by an active-layer detachment causes this type of slide to develop. Photograph by L.D. Dyke. GSC 2000-035C*

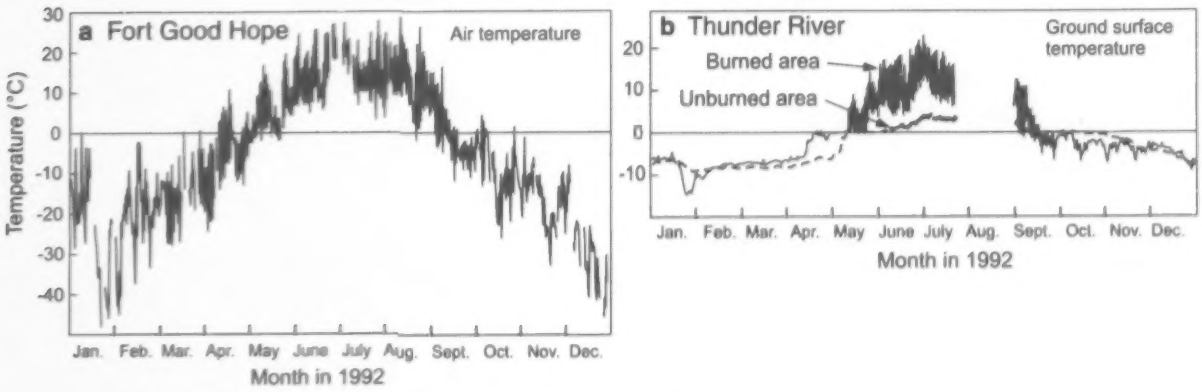

Figure 10. a) *Air temperature record for Fort Good Hope in 1992 (170 km southeast of the Thunder River site) and* ***b)*** *simultaneous temperature records for the ground surface in burned and unburned areas at the Thunder River site. The ground surface receives about four times the thawing degree days received at the unburned site.*

the burned area, destruction of the organic cover was not complete, with a charred mat of mosses remaining. The unburned site is characterized by a moss, lichen, and dwarf shrub mat. Of course, the contrast in temperature between burned and unburned sites will vary depending on the kind and thickness of original vegetation cover and the degree of destruction. At Thunder River, the ground surface at the burned site is receiving about four times the degree-days of thawing received at the unburned site. A comparison of seasonal thaw calculated for these two environments is shown in Figure 11. This graph shows the progress of thawing based on the average temperature during the thaw season. Although the maximum thaw depth reached is accurate, the progress of thaw until this depth is reached will probably be different from that indicated on the graph. The rate of thaw will be determined by the temperature variability during the thaw season and the amount of ground ice present at different depths.

Both the increase in thaw depth and more rapid melting of ground ice on a burned slope will favour active-layer detachments. More rapid thaw will reduce friction and deeper thaw will increase the weight of material available to overcome cohesion. The sensitivity of slope stability to changes in these factors is shown in Figure 12. It gives the maximum slope angle that will be stable, depending on the depth of thaw and the amount of friction afforded by the soil. The other component of soil strength, cohesion, was determined in the same way that the cohesion of the frozen sand at Mountain River was determined. An active-layer detachment with known slide thickness and slope angle was chosen and the cohesion required to just balance the force causing failure was calculated. This requires an assumption about the frictional component of strength to be made. Considering that ground ice is abundant in the tills of the area and these materials contain approximately 25% clay, the likelihood is great that enough meltwater could be retained within the soil to support the

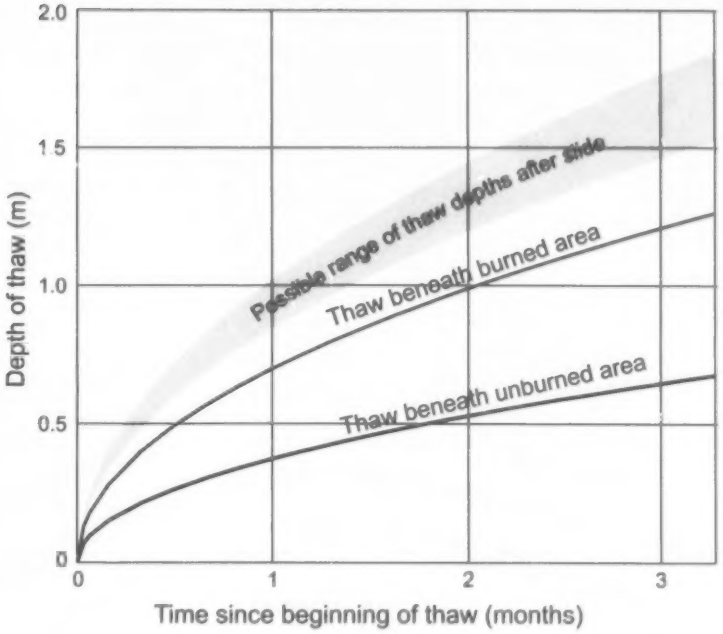

Figure 11.

Progress of frost table deepening under different ground-surface conditions for the Thunder River area. The curves are calculated using the average ground-surface temperature acting over the thawing interval for the burned and unburned site records shown in Figure 10. Burning of the ground cover doubles the active-layer thickness during the first thaw season. If an active-layer detachment occurs, an even deeper thaw is produced, assuming that the air temperature is an accurate estimate of the ground-surface temperature.

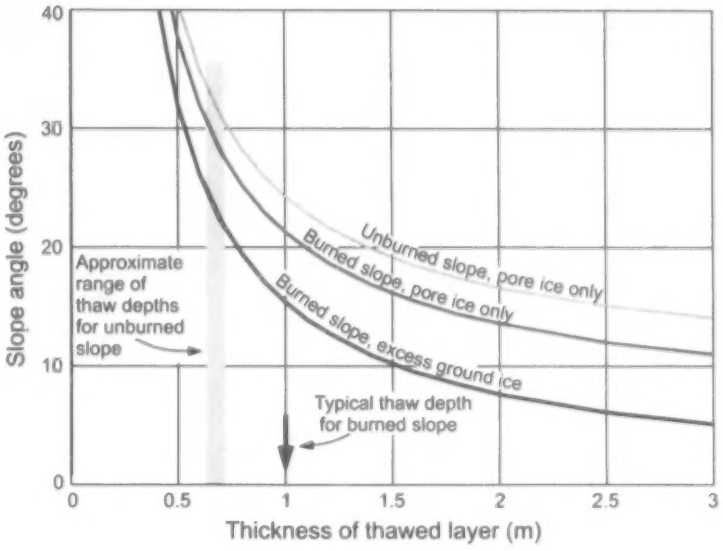

Figure 12.

Slope angles for which the factor of safety for an active-layer detachment = 1, depending on the depth of thaw and the amount of ground ice. The least stable slopes are burned slopes with ground ice in excess of that filling sediment pore space. Increases in average annual air temperature given by general circulation models alone will not have a pronounced affect on slope stability for this type of slide; however, the strong association with slope burning suggests that active-layer slides will become more common if climate change results in an increased incidence of forest fires.

entire thawed soil weight. Therefore, one assumption is that the frictional strength is zero (or $\phi = 0$). If ice in the soil is present only in the pore space, a higher friction will be available but thaw consolidation will still occur and still increase the pore-water pressure. Figure 12 shows examples of the effect of these different soil conditions.

Figure 12 indicates that an active-layer detachment will only form on steep slopes of 30° or more for thaw depths typical of undisturbed terrain in the Thunder River area (maximum thaw depths are 60–70 cm in the Thunder River area (Harry and MacInnes, 1988)). Active-layer detachments on undisturbed terrain in this area are rare, probably because slopes, except river banks, are seldom this steep; however, the maximum slope angle likely to fail decreases markedly as the thaw depth increases. With double the normal thaw depth, a much larger area becomes susceptible to failure because slopes as shallow as 10–15° become unstable. Of course, the likelihood of failure is still uncertain because the amount of ground ice is unknown. Ice characterized by visible lenses is most likely to eliminate friction because melting ice lenses are probably replaced by temporary lenses of water having no shear strength. Ice present only in existing sediment pore space will permit some friction because melting allows sediment particles to come into contact more rapidly. The volume and fabric of ice thus become important in determining stability.

SLOPE STABILITY IN A WARMING CLIMATE

Rotational slides in permafrost will become more likely if climate warming results in decreasing permafrost thickness and hence a decrease in the cohesive strength of sediments forming a slope; however, once a new permafrost thickness is established, slide triggering will revert to such processes as toe erosion. Permafrost thaw may be so gradual that failures due to the corresponding loss of strength may be difficult to distinguish from triggering by active toe erosion. The permafrost thaw may be a factor that simply hastens failure due to other destabilizing processes.

Climate warming may have a more noticeable effect on active-layer detachments; however, the warming must be pronounced for the effect to be noticeable. The gradual warming implied by General Circulation Model predictions amounts to a small fraction of a Celsius degree per year and would require years or decades to produce a marked increase in slide occurrence. Year-to-year variability is likely to be more effective in the short term. For example, the air thawing index averages 1600 degree days at Fort Good Hope but has been as high as about 1900 degree days. Therefore, in any one year, a thawing index considerably greater than the average is possible, resulting in an increase in thaw depth that would only be approached after many years of climate warming at predicted rates. As pointed out already, forest fire is probably the most likely means of triggering an active-layer detachment because of the marked affect on the transfer of heat into the ground. If climate warming is characterized by conditions which become more favourable for forest fire occurrence, then the frequency of active-layer detachments may increase. This may be the most pronounced influence of climate change on slope stability.

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Potential changes in thaw depth and thaw settlement for three locations in the Mackenzie valley

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Burgess, M.M., Desrochers, D.T., and Saunders, R., 2000: Potential changes in thaw depth and thaw settlement for three locations in the Mackenzie valley; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 187–195.*

Abstract: Locations near Tuktoyaktuk, Norman Wells, and Fort Simpson in the Mackenzie valley are analyzed to determine how quickly permafrost would degrade if the climate warms. The warming predicted by Canada Climate Centre modelling of climate under a doubled concentration of atmospheric carbon dioxide (5°C for Tuktoyaktuk, 4.2°C for both Norman Wells and Fort Simpson) was applied at different rates over a 50 year period. The effects of both and increase and decrease of snowfall were examined. After 50 years, the warming only produces an increase in active layer thickness at Tuktoyaktuk. At Norman Wells, thaw of permafrost to a depth approaching 10 m occurs for low ice content ground. At Fort Simpson, almost all permafrost disappears. Thaw subsidence associated with the modelled 50 year degradation of ice-rich permafrost ranges from 0.5 m at Norman Wells to more than 1.5 m in the Fort Simpson region.

Résumé : On a analysé des sites situés près de Tuktoyaktuk, Norman Wells et Fort Simpson dans la vallée du Mackenzie pour déterminer la rapidité à laquelle le pergélisol se dégraderait s'il y avait un réchauffement climatique. Le réchauffement prévu par le Centre canadien de climatologie, si la concentration de dioxyde de carbone atmosphérique devait doubler (5 °C pour Tuktoyaktuk; 4,2 °C pour Norman Wells et Fort Simpson), a été appliqué à différents taux sur une période de 50 ans. Étant donné que ce modèle climatique prévoit également une légère augmentation des précipitations, on a également analysé les effets d'une neige plus abondante. Après 50 ans, le réchauffement n'augmenterait l'épaisseur du mollisol qu'à Tuktoyaktuk. À Norman Wells, le dégel du pergélisol jusqu'à une profondeur pouvant atteindre 10 m aurait lieu dans les sols à faible teneur en glace. À Fort Simpson, le pergélisol disparaîtrait presque complètement. Selon le modèle de la dégénérescence d'un pergélisol riche en glace échelonnée sur 50 ans, la subsidence due au dégel varierait de 0,5 m à Norman Wells à plus de 1,5 m à Fort Simpson.

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INTRODUCTION

This paper discusses how climate warming due to doubling atmospheric carbon dioxide in 50 years may affect permafrost at specific locations in the Mackenzie valley. Three locations in the Mackenzie valley are examined: Richards Island near Tuktoyaktuk, Norman Wells, and Fort Simpson (Fig. 1). At each of these, an ice-poor and an ice-rich site are considered. The response of temperature in the ground to temperature in the air is predicted using a mathematical simulation of heat flow between the air and the ground (geothermal model). These simulations apply a prescribed change in air temperature over the 50 year period and show how ground temperatures change and permafrost thickness decreases. The ground ice content is an important factor in determining how quickly permafrost disappears. The amount of ice melted is then combined with data on soil type to calculate potential thaw settlement. Performing simulations for different initial climates (i.e. different regions), different soil conditions, and different degrees of climate change provides some indication of the potential range in permafrost response throughout the Mackenzie valley.

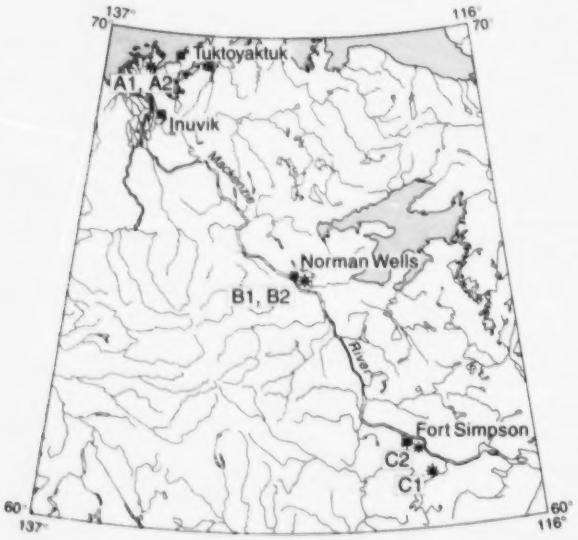

Figure 1. *Location of study sites. Permafrost boundaries are* modified from *Heginbottom et al. (1995).*

To carry out the simulation, detailed information is required for each site: borehole data (sediment type, physical and thermal properties), ground temperature measurements, surface condition, ice content, and climate data (air temperature and precipitation). There is no unique method to undertaking this modelling and the approach used here is briefly outlined. A more detailed discussion of procedures can be found in GSC Open File 3017 (Geo-Engineering (M.S.T.) Ltd., 1994).

CLIMATE DATA

The 1951–1980 climate normals for mean monthly air temperature and snow cover at the nearest Atmospheric Environment Service (AES) climate stations, i.e. Tuktoyaktuk, Norman Wells, and Fort Simpson, were adopted as the initial climate conditions corresponding to the present day carbon dioxide CO_2 concentration (i.e. 1 x CO_2 conditions). Atmospheric Environment Service's General Circulation Model (GCM) of mean monthly air temperature and mean monthly total precipitation under a doubled CO_2 concentration (i.e. 2 x CO_2 conditions) were then obtained (Atmospheric Environment Service, pers. comm., 1992) for each of these stations. Table 1 summarizes the present and predicted annual climate conditions for each site. The change from 1 x CO_2 to 2 x CO_2 conditions represents a 4°C to 5°C increase in mean annual air temperature, and increases of 9–16% in total annual precipitation.

Ground temperature in permafrost regions is greatly influenced by thickness and duration of snow cover (Goodrich, 1982). A discussion of observed trends in air temperature and snow depths at climate stations in the Mackenzie valley during the 20th century and their relationship to permafrost stability is presented in Stuart et al. (1991). Snow cover changes associated with the changing climate should therefore be incorporated in the geothermal simulation. Since the 2 x CO_2 climate prediction provided only total water-equivalent precipitation data, a means to estimate future snowfall and snow depth from the monthly precipitation data was therefore required. A method was developed based on the current relationship between temperature and snowfall. The fraction of total precipitation that falls as snow for each month, based on climate data from 1951–1980, is related to the monthly mean temperature for each Atmospheric Environment Service station (Fig. 2a shows the result for Norman Wells). This links snowfall to available parameters that are predicted by the climate change model. With Figure 2a, the amount of future snowfall can be estimated using the new mean monthly temperature provided by the climate model. To determine the

Table 1. Summary of mean annual climate conditions at start of simulation (1951–1980 climate normals) and end of climate warming (2 x CO_2 conditions).

	MEAN ANNUAL AIR TEMPERATURE (°C)			TOTAL ANNUAL PRECIPITATION (mm)		
LOCATION	30 year normals (1951–1980)	2XCO₂ scenario	Difference	30 year normals (1951–1980)	2XCO₂ scenario	Difference
Tuktoyaktuk	-10.9	-5.9	+5.0	138	157	+19
Norman Wells	-6.4	-2.2	+4.2	328	356	+28
Fort Simpson	-4.2	-0.0	+4.2	355	411	+56

actual depth of snow throughout the year, an additional step to relate snowfall to depth of snow on the ground for each month that snow accumulates or ablates is necessary. This is done in Figure 2b for accumulation and Figure 2c for ablation; these figures are also based on 1951–1980 climate data. With these correlations, snowpack history under the changed climate can be constructed.

Plots of 1951–1980 monthly climate data and the 2 x CO_2 monthly air temperatures and derived mean monthly snow depths are given for each region in Figure 3. Note that at all three locations, although there is an increase in total annual precipitation under 2 x CO_2 conditions, mean monthly snow depths have decreased under the warming scenario. This is due in part to warmer air temperatures causing more precipitation to fall as rain rather than snow. For the purpose of modelling the response of permafrost, these changes in mean monthly air temperature and snow cover are allowed to occur over the 50 year time interval during which the CO_2 doubling is assumed to take place. Figure 4 shows possible paths to reach the 2 x CO_2 conditions that were examined in the simulations: 1) linear case with a linear increase in temperature and linear decrease in snow cover, and 2) exponential increase in temperature and exponential decrease in snow depth. Since there is considerable uncertainty in climate model precipitation predictions and how these are translated into snow cover estimates, an additional scenario was included; 3) a linear increase in temperature with a 10% linear increase in snow cover (subsequently referred to as linear+10%). Scenario 3 thus allows an examination of the sensitivity of permafrost to an increase in snow cover.

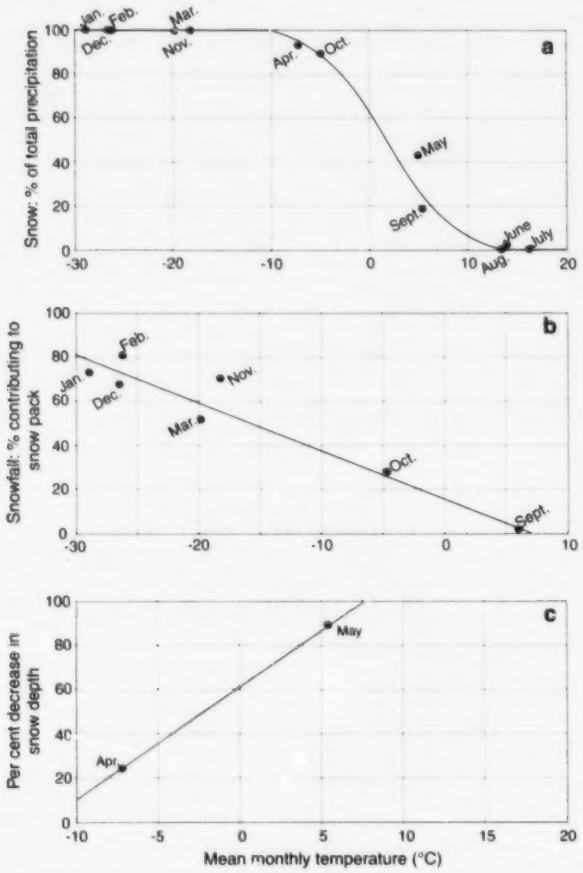

Figure 2. *Correlations established for Norman Wells climate data to allow conversion of precipitation predicted by climate change models into snowfall and snow cover, based on monthly mean air temperature and precipitation data from the 1951–1980 climate data.* ***a)*** *Percentage of total precipitation that falls as snow for each month as a function of mean monthly temperature.* ***b)*** *Percentage of monthly snowfall that accumulates in the snowpack, i.e. conversion factor for snowfall into depth of snow on the ground, as a function of mean monthly temperature.* ***c)*** *Ablation rate of the snowpack in spring as a function of mean monthly temperature.*

SITE SELECTION

The three study areas were selected to provide a transect from cold continuous permafrost along the Beaufort Sea coast to warm discontinuous permafrost near Fort Simpson. Area A is located along a borehole transect established by the GSC to study massive ground ice in the Mackenzie Delta. At site A1 a massive ice layer underlies a thin glaciofluvial layer of silt and sand. Site A2 represents a relatively ice-poor glaciofluvial deposit. Average ground surface temperatures are -6 to -7°C at these sites and permafrost is several hundred metres deep in this region (*see* Taylor et al., 2000).

Sites in area B, near Norman Wells, were selected from numerous geotechnical boreholes drilled for design purposes by Interprovincial Pipeline Inc. (IPL), and by both IPL and the federal government for permafrost research along the Norman Wells pipeline corridor. Site B1 is located near Norman Wells on an ice-rich lacustrine plain (IPL borehole 80-1). Site B2 is located 19 km south of Norman Wells along the pipeline, near Canyon Creek, on a low-ice content till plain (pipeline study site 84-2A; Pilon et al., 1989). Permafrost is 25–50 m thick at these sites and mean annual ground surface temperatures are -1°C to -2°C.

Area C is located in the discontinuous permafrost region of Fort Simpson. The two sites here are approximately 50 km apart. Both are located along the Norman Wells pipeline and have very detailed geotechnical and geothermal data (Patterson and Riseborough, 1988; Patterson et al., 1988; Pilon et al., 1989). For a discussion of other aspects of Norman Wells Pipeline permafrost and terrain data, *see* Burgess and Lawrence (2000). Site C1 (pipeline study site 85-12B) is situated in an ice-rich peat plateau; permafrost is around 5.6 m thick. Site C2 is ice-poor in the near surface, but has an ice-rich clay at depth; permafrost is approximately 12 m thick. At both sites the mean annual ground surface temperature is between -1°C and 0°C.

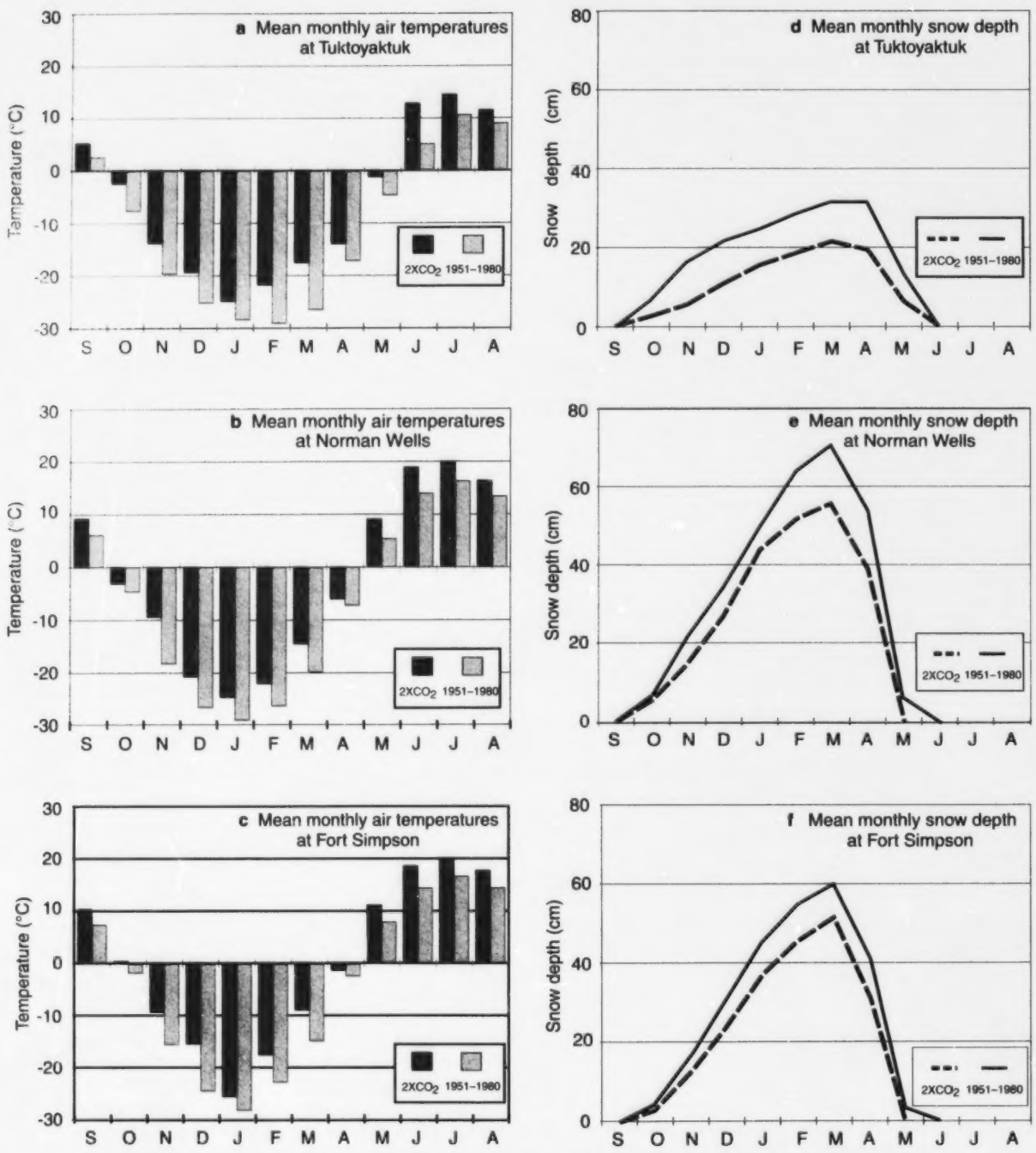
a Mean monthly air temperatures at Tuktoyaktuk
b Mean monthly air temperatures at Norman Wells
c Mean monthly air temperatures at Fort Simpson
d Mean monthly snow depth at Tuktoyaktuk
e Mean monthly snow depth at Norman Wells
f Mean monthly snow depth at Fort Simpson
Temperature (°C)
Snow depth (cm)
2XCO2 1951–1980
S O N D J F M A M J J A

Figure 3. a), b), c) Mean monthly air temperatures and d), e), f) snow depths from 1951–1980 climate normals and predicted 2 x CO_2 conditions at the Tuktoyaktuk, Norman Wells, and Fort Simpson climate stations, respectively.

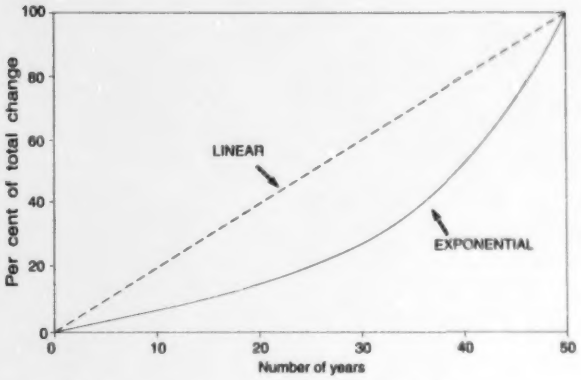

Figure 4. *Paths of change used to reach the 2 x CO_2 climate over a 50 year interval. The y-axis represents the per cent change, with 1 x CO_2 conditions equal to 0% and 2 x CO_2 conditions equal to 100%.*

Figure 5 presents for each borehole, logs of the lithology, moisture content or ice content, and temperature envelope (annual minimum and maximum) used in the geothermal simulations. Thaw strains, also included in these figures, will be discussed separately in "Potential thaw settlement" section.

GEOTHERMAL MODEL

The geothermal analyses were performed using a computer model (the one-dimensional model described in Nixon (1983)) that predicts depth of thawing using air temperature and snow depth data (i.e. the climate histories developed above). The model requires soil moisture content, soil density, and frozen and thawed thermal conductivity, determined when possible from borehole samples. Details on procedures to ensure that the model's initial ground thermal conditions were stable (i.e. in equilibrium with the model's initial climate conditions) are provided in GSC Open File 3017. Air temperatures were applied to the top of the snowpack in winter and to the ground surface in summer. An empirical relationship between air temperature and the actual temperature experienced at the ground surface (the n-factor, *see* Taylor (2000)) was used to account for the attenuation of air temperature caused by surface-layer conditions such as vegetation and snow.

THAW-DEPTH PREDICTIONS

Predicted thaw depths are plotted as a function of time for the three climate change regimes (linear, exponential, and linear+10%) in Figure 6. These plots show the rate of increase in maximum annual thaw depth over the 50 year simulation period. At the Beaufort Sea coast sites, only small increases in active layer (<0.5 m) are observed, regardless of the scenario, with the smallest increase occurring at the ice-rich site. The bulk of the energy reaching the ground during the 50 year climate warming in this region is directed towards increasing the permafrost ground temperatures (as shown for site A1 in Fig. 7), rather than in melting ground ice.

At Norman Wells, increases in thaw depth for the ice-rich site (B1) are very slight for the linear and exponential case, 0.5 m and 0.7 m respectively, but about 2.5 m for the linear+10% case. For the low-ice site B2, increases in thaw depth are much larger for all scenarios, ranging from about 2 m for the linear case to almost 7 m for the linear+10% case. The rates of increase in thaw depth are less at the ice-rich site because of the latent heat required to melt the additional ground ice. After 50 years of warming, thaw depths could range from about 1.5 m to 8 m in this region depending on warming scenario and soil profile. With time, allowing ground temperatures to return to equilibrium under the 2 x CO_2 conditions, permafrost would likely completely disappear at these sites, as can be deduced from an examination of the ground-temperature profiles shown for site B2 in Figure 7. Norman Wells permafrost occurrence would resemble conditions presently encountered farther south in the sporadic zone.

At Fort Simpson, changes in thaw depth are even more dramatic and complete degradation of permafrost may well occur in 50 years or less. At the peat plateau site C1, where permafrost initially was 5.6 m thick, both the linear and linear+10% scenarios caused permafrost to degrade in less than 50 years. At site C2, the 12 m of permafrost would degrade in about 50 years under the linear +10% scenario, and likely within 100 years for the other two scenarios. Of note, the rate of retreat of the base of permafrost is slower in the C2 case due to the presence of higher ice content at depth at this site.

It is interesting to note that in all simulations for Norman Wells and Fort Simpson (where impacts on thaw depths are the greatest), the rate of change in thaw depth is very slow in the first 10–20 years of the simulations. This observation suggests that detection of the early impacts of warming by monitoring thaw depth (*see* Nixon, 2000) may be difficult, particularly when allowing for seasonal variations and when dealing with permafrost temperatures near 0°C where heat associated with phase change plays an important role in the ground thermal regime (Burgess and Riseborough, 1990; Riseborough, 1990). The simulations undertaken here support the need for long-term observations to clearly define trends.

POTENTIAL THAW SETTLEMENT

The potential for settlement of the ground surface as frozen ground thaws at each site is determined using the calculated thaw strain for each soil layer. The total thaw strain is a measure of the per cent change in height or thickness of a soil unit upon thaw (i.e. per cent settlement), and is a function not only of the volume change involved in melting frozen ground but also of the compressibility of the soil layer. The compressibility component is generally one to two orders of magnitude smaller than that due to thaw, particularly for the thin layer (less than 10 m) of overlying material under consideration here, and therefore it has not been included in the

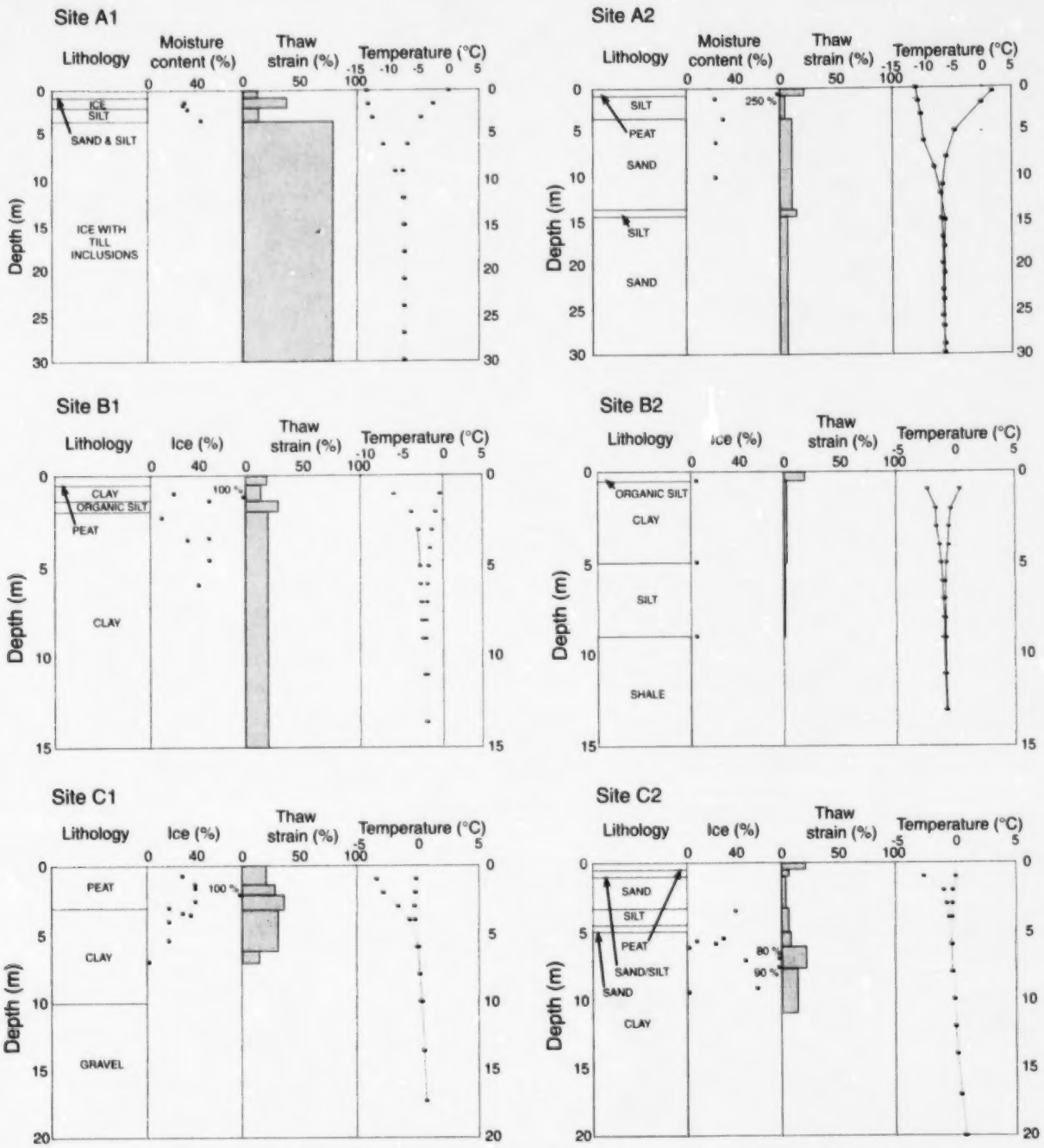

Figure 5. *Stratigraphy, moisture content, thaw strain in the case of complete melting of ground ice, and ground temperature envelopes for each borehole used in the study.*

calculations. Thaw relationships reported in the literature for major soil groups (Speer et al., 1973; McRoberts et al., 1978a, b; Hardy Associates Ltd.,, 1982;) were used to determine thaw strain. These require a knowledge of sediment type, and of one of the following properties: total moisture content, ice content, or frozen bulk density. Generally the geotechnical property most readily available for these determinations is the moisture content. Calculated thaw-strain values for each soil layer are included in the borehole data plotted in Figure 5.

The calculated thaw strains were then used to determine the thaw settlement resulting from the simulated increase in thaw depth at each site for each scenario. Thaw settlements after 10 and 50 years are tabulated in Table 2. Sites and scenarios which have complete permafrost degradation over the 50 years are shaded in the table and the predicted settlements for these cases are the ultimate settlement for the site. Elsewhere, since complete permafrost degradation has not occurred during the 50 years, the potential for additional thaw settlement obviously remains.

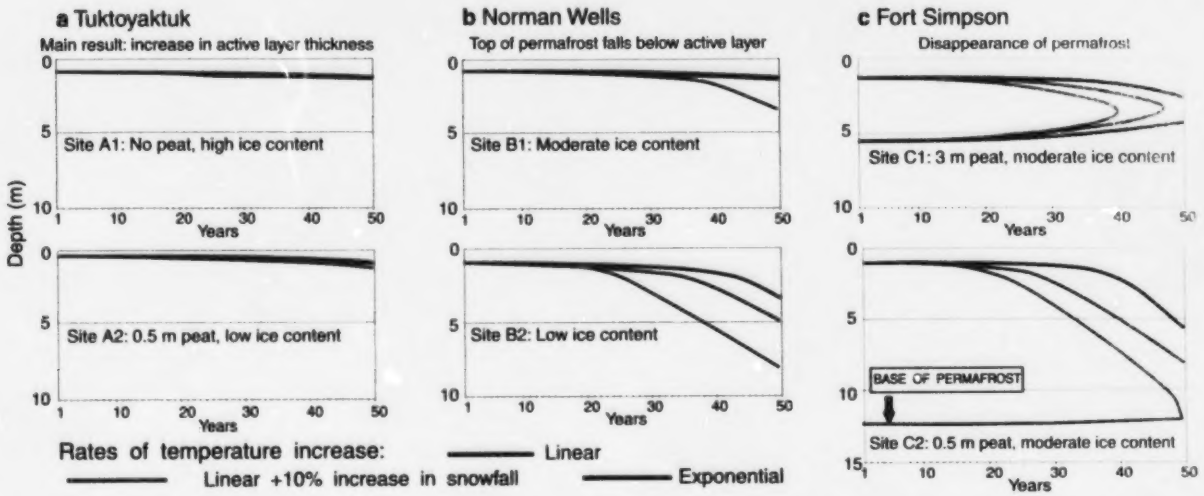
a Tuktoyaktuk
Main result: increase in active layer thickness
Site A1: No peat, high ice content
Site A2: 0.5 m peat, low ice content
b Norman Wells
Top of permafrost falls below active layer
Site B1: Moderate ice content
Site B2: Low ice content
c Fort Simpson
Disappearance of permafrost
Site C1: 3 m peat, moderate ice content
BASE OF PERMAFROST
Site C2: 0.5 m peat, moderate ice content
Depth (m)
Years
Rates of temperature increase:
Linear
Linear +10% increase in snowfall
Exponential

Figure 6. *Predicted maximum annual thaw depth for each site, under linear, exponential, and linear+10% temperature change regimes.*

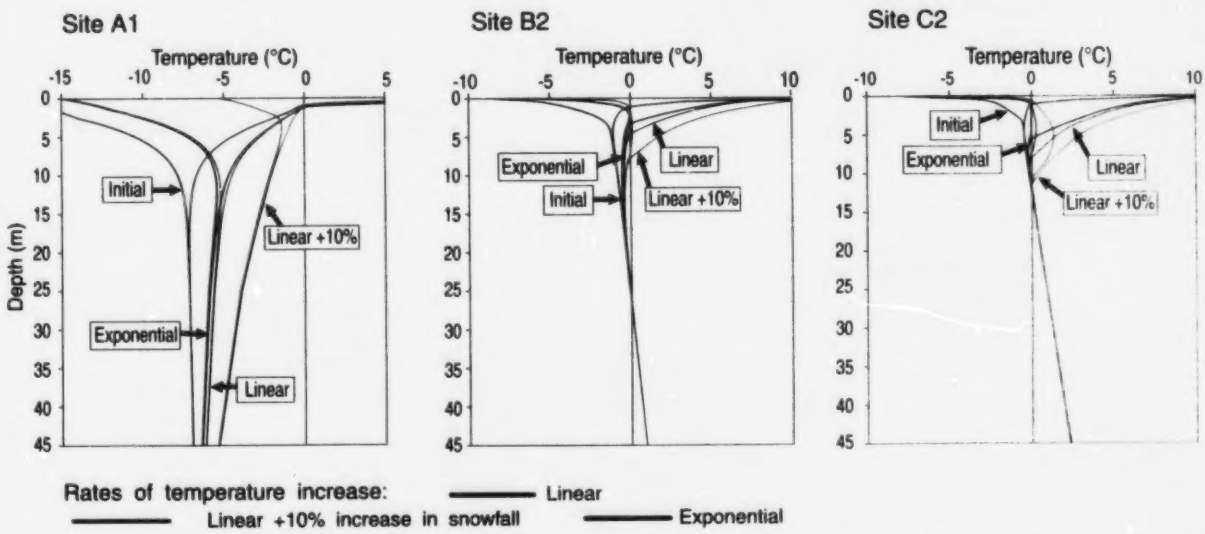

Figure 7. *Ground temperature profiles predicted after 50 years for site A1, B2, and C2 under the temperature change regimes shown in Figure 5. Profiles used as the start-up conditions at these sites are also shown (initial case).*

Table 2. Predicted thaw depths and calculated thaw settlements after 10 years and 50 years at each site for each climate-change history. Shaded areas indicate that complete permafrost degradation and maximum thaw settlement have occurred.

Site	Surface function	Active layer thickness (m)		Cumulative thaw settlement (m)	
		t=10 years	t=50 years	t=10 years	t=50 years
A1	Exponential	0.9	1.2	0.00	0.12
	Linear	1.0	1.2	0.02	0.12
	Linear +10%	1.0	1.3	0.02	0.16
A2	Exponential	0.4	0.6	0.00	0.03
	Linear	0.4	0.6	0.00	0.03
	Linear +10%	0.4	0.7	0.00	0.06
B1	Exponential	0.9	1.3	0.00	0.12
	Linear	1.0	1.4	0.03	0.14
	Linear +10%	1.0	3.1	0.03	0.50
B2	Exponential	1.1	3.5	0.00	0.01
	Linear	1.3	4.7	0.04	0.06
	Linear +10%	1.3	7.8	0.04	0.08
C1	Exponential	0.9	2.3	0.00	0.38
	Linear	1.0		0.01	
	Linear +10%	1.0		0.01	
C2	Exponential	1.1	6.0	0.00	0.50
	Linear	1.2	8.0	0.01	1.06
	Linear +10%	1.2		0.01	

At the Beaufort Sea sites, where increases in thaw depths are small, thaw settlement is at most a few centimetres. In the Norman Wells region, after 50 years, thaw settlements range from less than 0.15 m for all scenarios at the low-ice site B2 to 0.5 m at the ice-rich site B1 for the linear+10% case. At Fort Simpson, where permafrost degrades completely at site C1 in less than 50 years for all cases except the exponential, this results in 1.44 m of settlement. At site C2, permafrost degrades completely in 50 years or less for the step and linear+10% cases, resulting in about 1.8 m of settlement. The exponential case for site C2, which produced the slowest increase in thaw depth, still results in 0.5 m of settlement.

In summary for the sites and approach selected for this analysis, thaw settlement over 50 years is negligible in the Tuktoyaktuk area, and could range from 0.01 m to 0.5 m in the Norman Wells region and from 0.4 m to 1.8 m in the Fort Simpson area.

The results of the modelling exercise presented in this chapter clearly indicate how permafrost response to climate warming can be expected to vary, depending on site-specific characteristics and the climate change history. A modelling approach that takes into account normal year-to-year climatic variability along with the predicted warming in the Fort Simpson region can be found in Riseborough and Smith (1993). For a discussion of mapping the response of permafrost to climate warming, *see* Wright et al. (2000).

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Potential changes in permafrost distribution in the Fort Simpson and Norman Wells regions

J.F. Wright[1], M.W. Smith[2], and A.E. Taylor[3]

Wright, J.F., Smith, M.W., and Taylor, A.E., 2000: Potential changes in permafrost distribution in the Fort Simpson and Norman Wells regions; in *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547, p. 197–207.*

Abstract: Modelling that predicts permafrost distribution and thickness in equilibrium with a given mean annual air temperature is applied to the Norman Wells and Fort Simpson regions. The model predicts the likelihood of permafrost, based on values of thermal conductivity for the various surficial materials in each study area and a factor which describes the insulating property of ground-surface vegetation and snow cover. Estimates of permafrost thickness are obtained for various combinations of terrain characteristics. Using maps of vegetation and surficial geology, these combinations can be compiled for each study area and used to map both permafrost thickness and extent, using a geographic information system. This technique predicts that, under an increase in mean annual air temperature of 2°C, permafrost extent decreases slightly and thickness decreases markedly for the Norman Wells area. For the same temperature increase at Fort Simpson, permafrost almost completely disappears.

Résumé : On a appliqué un modèle prévoyant un équilibre entre la répartition et l'épaisseur du pergélisol et la température annuelle moyenne de l'air à des secteurs de 40 km^2 situés près de Norman Wells et de Fort Simpson. Le modèle calcule la vraisemblance d'un pergélisol, en se fondant sur la conductivité thermique des différents matériaux superficiels dont est composé chaque secteur à l'étude ainsi que sur un facteur décrivant les propriétés d'isolation de la végétation et du manteau nival. On obtient des estimations de l'épaisseur du pergélisol dans des terrains affichant des combinaisons de caractéristiques variées. En utilisant les cartes de la végétation et des dépôts superficiels, on peut compiler ces combinaisons et cartographier l'épaisseur et l'étendue du pergélisol dans chaque secteur à l'étude au moyen d'un système d'information géographique. Cette technique permet de prévoir que, si la température annuelle moyenne de l'air devait augmenter de 2 °C, l'étendue du pergélisol diminuerait légèrement et son épaisseur fléchirait encore plus dans la région de Norman Wells. Pour une augmentation semblable de la température à Fort Simpson, le pergélisol disparaîtrait presque complètement.

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INTRODUCTION

Modelling the response of climate to increased atmospheric CO_2 concentrations indicates that global temperatures will continue to warm gradually. This paper describes the predicted impact of climate warming upon the occurrence and thickness of permafrost at Fort Simpson and Norman Wells, Northwest Territories. The modelling follows a technique developed by Jorgenson and Kreig (1988), which incorporates key environmental factors for predicting the occurrence of permafrost. The Jorgenson and Kreig (J&K) model has been expanded to derive a 'Hybrid Permafrost Prediction Model' (HPPM) that estimates permafrost thickness as well as distribution. It has been applied to the Fort Simpson and Norman Wells areas (Wright, 1995). The model is implemented within a Geographical Information System (GIS) to predict the equilibrium distribution of permafrost at the two study sites under present climatic conditions and under increases in mean annual air temperature of 1°C and 2°C.

Fort Simpson study area

The Fort Simpson study area surrounds the confluence of the Mackenzie and Liard rivers in the Northwest Territories (Fig. 1). It is situated within the zone of discontinuous permafrost and thus provides a good test location for evaluating the ability of permafrost models to differentiate between frozen and unfrozen ground. The surficial geology is characterized by extensive glaciolacustrine deposits in the lowlands south of the Mackenzie River and west of the Liard River (Fig. 2). A considerable portion of these lowlands is overlain by organic veneers associated with bogs and fens. North of the Mackenzie River a broad till plain is present. On either side of the Liard River, eolian sands create an undulating dune topography,

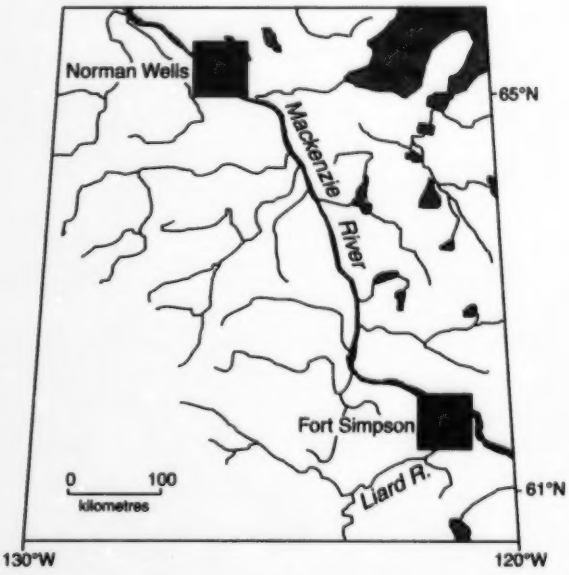


Figure 1. *Locations map for the Fort Simpson and Norman Wells study areas.*

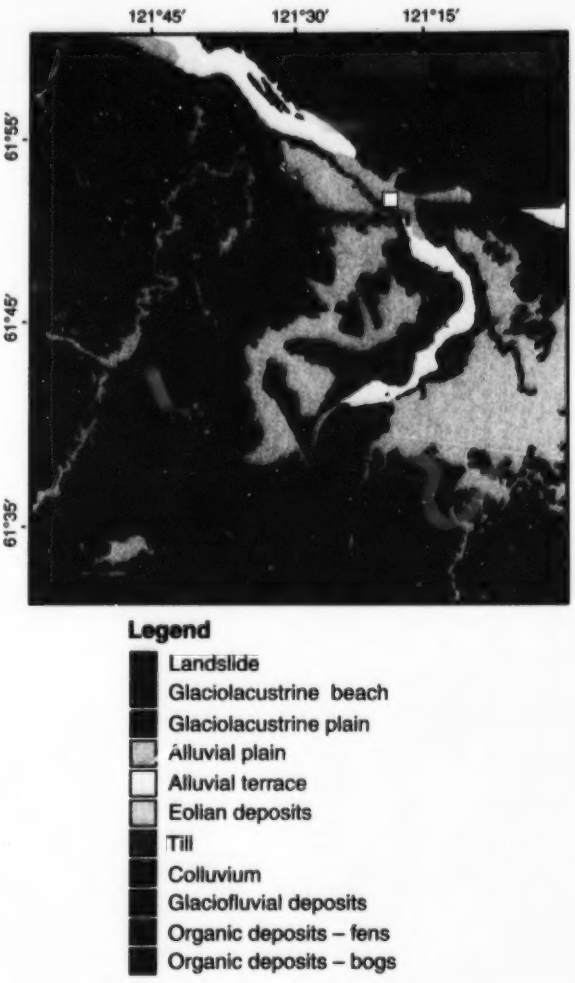
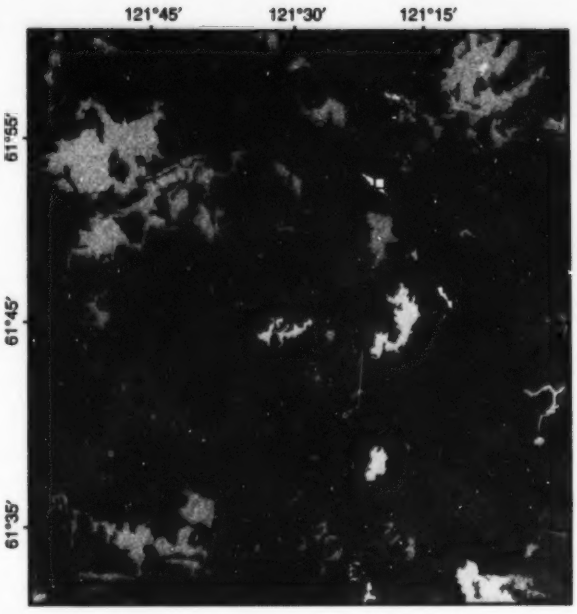


Figure 2. *Surficial geology of the Fort Simpson study area.*

with organic deposits filling interdune depressions. An extensive fenland covers much of the southeast corner of the study area. Regional vegetation characteristics reflect the generally poor drainage associated with the subdued topography (Fig. 3), with black spruce forests, bogs, and fenland dominating the lowlands. White spruce forests occupy moderately well drained sites, whereas upland pine and deciduous species are found in well drained environments. Thirty-year climate normals (1961–1990) for Fort Simpson indicate a mean annual air temperature of -4°C, with an annual variation in mean monthly temperature of 23°C.

Norman Wells study area

A second study area of similar size is located at Norman Wells, approximately 400 km north of Fort Simpson. The Norman Wells area is characterized by widespread permafrost. This affords an opportunity to evaluate Hybrid Permafrost Prediction Model performance in a location that



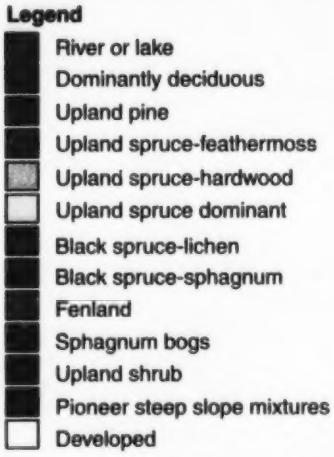


Figure 3. *Vegetation in the Fort Simpson study area.*

experiences a significantly cooler climate than that at Fort Simpson. Vegetation and soil types are similar, but the Norman Wells region is dominated by a broad expanse of black spruce forest underlain by glaciolacustrine silt deposits. Organic units such as bogs and fens make up a very small proportion of the study area, compared with their relative dominance at Fort Simpson (Fig. 4). Deciduous forests are also rare, and pine trees are virtually absent. The Franklin Range of mountains parallels the Mackenzie River to the north. Most of this range has been excluded from this analysis, as has a section of the Mackenzie Mountains in the southwest corner of the study area. Climate normals for Norman Wells indicate a mean annual air temperature of approximately -6°C and an annual amplitude of 23°C. Both the Fort Simpson and Norman Wells study areas are traversed by sections of the Norman Wells pipeline right-of-way (MacInnes et al., 1989).

AN INDEX OF PERMAFROST OCCURRENCE

Jorgenson and Kreig (1988) devised a model for predicting the occurrence of permafrost based on simple relations for estimating thawing and freezing depths under constant surface temperature. The ground thermal state (frozen or unfrozen) is assumed to be dependent on the relative duration and magnitude of air temperature above and below 0°C (expressed in degree-days), seasonal ground thermal

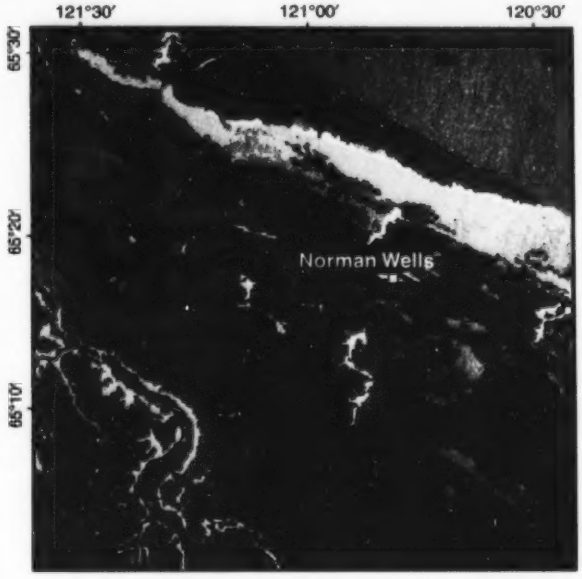


Figure 4. *Vegetation in the Norman Wells study area.*

conductivity (frozen and unfrozen), and heat transfer between the air and the ground, through intervening vegetation and snow cover. A permafrost index (R) is related to the climate and terrain factors as follows:

$$R = P \bullet \frac{DDT}{DDF} \bullet \frac{n_t}{n_f} \bullet \frac{Kt}{Kf}$$

where:

P = potential insolation index
DDT = thawing degree days
DDF = freezing degree days
$\mathbf{n_t}$= summer n-factor
$\mathbf{n_f}$= winter n-factor
Kt= unfrozen thermal conductivity
Kf= frozen thermal conductivity

Values for R that are less than 1 imply that the depth of winter ground freezing exceeds the depth of summer thaw, and thus permafrost is likely to occur. The smaller the value of R (for R < 1), the greater the likelihood that permafrost is present.

The ease with which changes in atmospheric temperatures may be transferred through vegetation or snow cover (i.e. across the 'buffer' layer) to the ground surface may be expressed as a seasonal 'n-factor' (Lunardini, 1978, 1981; *see also* Taylor, 2000). An n-factor of 1 indicates no resistance to heat transfer, while an n-factor of 0 suggests a perfect insulator. The n-factors therefore convert the air degree-day totals to estimates of degree-day values actually occurring at the ground surface. A 'potential insolation index' (P) accounts for extra heating on south-facing slopes and reduced heating on north-facing slopes due to, respectively, increased or decreased exposure to the sun during the summer season; P is defined as the ratio of the site latitude to the equivalent latitude calculated according to Lee (1962, 1964).

A METHOD FOR ESTIMATING PERMAFROST THICKNESS

According to Jorgenson and Kreig (1988), the value R=1 is defined as the threshold of permafrost occurrence. It seems reasonable to expect that values of R progressively less than 1 should be indicative of thicker permafrost. A hybrid model was developed in which values of R generated by the Jorgenson and Kreig equation are related to predictions of permafrost thickness produced by a finite-element heat conduction model called TONE (L.E. Goodrich, pers. comm., 1993). The finite-element approach to solving heat conduction problems is well established (Myers, 1971; Goodrich, 1982), and TONE has been used as a tool for investigating permafrost responses to climate change (Riseborough and Smith, 1993); however, finite-element models are complex, involving labourious calculations, and are not compatible with geographic information systems. Because TONE can utilize the same data on climate and site conditions as the Jorgenson and Kreig model, a prediction of permafrost thickness under conditions of thermal equilibrium and an associated value of R may be obtained for any specified set of model inputs. In order to develop a GIS-compatible model capable of predicting permafrost thickness, outputs from the Jorgenson and Kreig and TONE models were correlated to derive the Hybrid Permafrost Prediction Model, which produces estimates of equilibrium permafrost thickness comparable to TONE predictions, using much simpler calculations. Input values were chosen so as to represent a wide range of hypothetical climate and terrain conditions expected to occur in natural settings.

Jorgenson and Kreig's R versus TONE predictions of permafrost thickness

In this paper, the term 'permafrost thickness' refers to the distance from the ground surface to the base of the permafrost, irrespective of active-layer thickness, under conditions of a periodic steady state. Model runs were designed to generate predictions for a broad range of hypothetical terrain conditions to reveal a relation between TONE predictions of permafrost thickness and the associated values of R. Figure 5 plots TONE predictions for permafrost thickness against the calculated value of R for these hypothetical terrain conditions.

Permafrost was predicted to occur only when R values were less than 1, confirming the threshold value specified by Jorgenson and Kreig (1988). It is clear that, while the R index appears to be a good indicator of permafrost occurrence, it is not a very precise indicator of permafrost thickness. It is necessary, therefore, to identify the separate influences of the various input parameters to describe the manner in which they combine to produce thicker or thinner permafrost.

Isolating the influence of model parameters

When TONE predictions of permafrost thickness were plotted against the negative log of R, a definite pattern emerged, with the data for each terrain condition (i.e. a particular set of n-factor and thermal conductivity values) displaying a linear increase

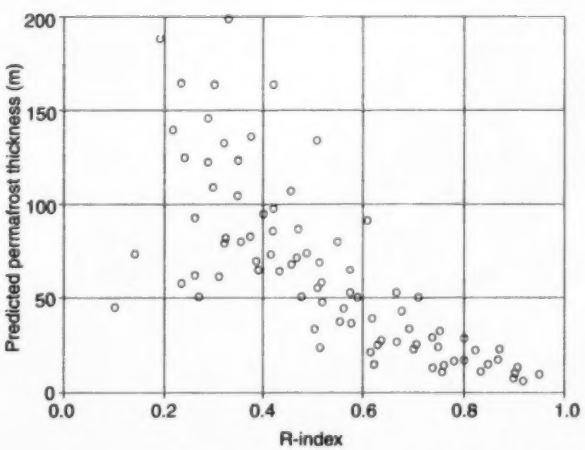


Figure 5. *Relations between the calculated value of R and permafrost thickness. Note that permafrost occurs only where R<1, but that R is a poor indicator of permafrost thickness.*

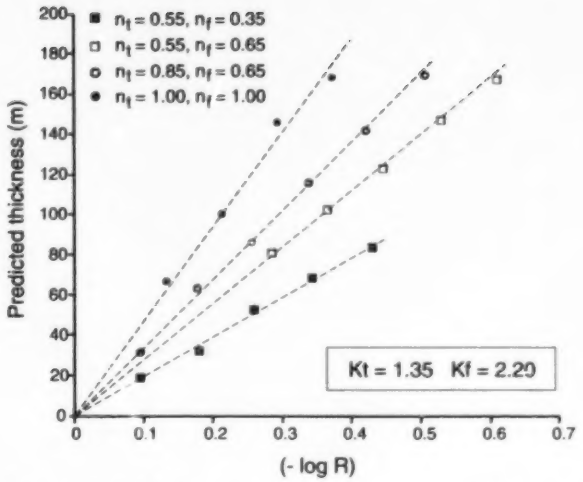


Figure 6. *Relations between R and TONE predictions of permafrost thickness. Note that all n-factor combinations produce slopes less than that at $n_t=n_f=1.0$, regardless of which n-factor is larger.*

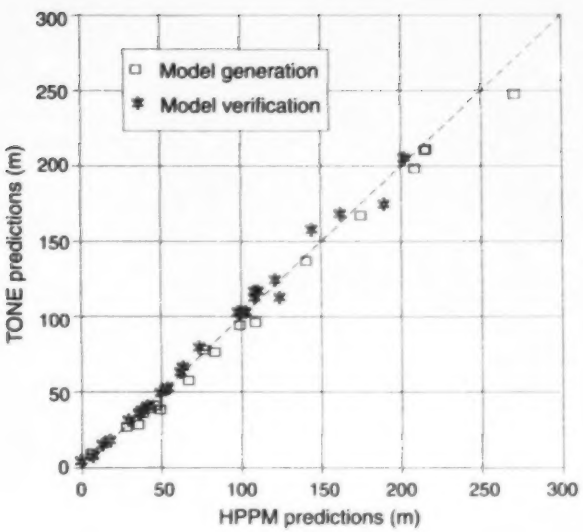


Figure 7. *Comparison of Hybrid Permafrost Prediction Model (HPPM) predictions of permafrost thickness to those generated by a finite-element heat conduction model (TONE).*

in TONE predictions for permafrost thickness as the value for -logR increases. Figure 6 displays data for four different n-factor pairs, given a single pair of seasonal thermal conductivity values. Since these data converge at the graph origin (0,0) the relations between permafrost thickness and -logR may be described for each terrain condition simply in terms of the slope (S) of the appropriate line. Data points along an individual line represent different mean annual temperature conditions at a particular hypothetical site (i.e. a single specified set of seasonal n-factor and thermal conductivity values). Since movement along a particular line represents a change in mean annual air temperature only, it follows that a change in mean annual air temperature does not influence the value of S. This implies that if an appropriate slope value (S) is determined for any specified set of terrain parameters (n_t, n_f, Kt, Kf), then the associated TONE prediction of permafrost thickness may be obtained directly, without the need for additional TONE runs.

Technical performance of the Hybrid Permafrost Prediction Model

The technical performance of the Hybrid Permafrost Prediction Model was evaluated with respect to its ability to reproduce the predictions of permafrost thickness generated by the finite-element heat conduction model (TONE). The evaluation involved 20 runs used in defining the mathematical relations comprising the Hybrid Permafrost Prediction Model, plus 50 additional verification runs whose parameters were chosen in a more-or-less random fashion. Figure 7 shows the comparison of TONE outputs to Hybrid Permafrost Prediction Model predictions of permafrost thickness for these 70 distinct terrain conditions. Predictions of permafrost thickness ranged from a few metres to nearly 300 m.

Linear regression of TONE estimates of permafrost thickness against Hybrid Permafrost Prediction Model predictions indicate a standard error across the whole range of predictions of 7.4 m, with an r^2 of 0.99. For predictions less than 50 m, the standard error is only 4.1 m (r^2 = 0.93), while for those between 50 m and 300 m the standard error is 9.4 m with an r^2 of 0.98. The standard error falls to 2.0 m for predictions less than 30 m, with an r^2 of 0.97. This analysis indicates that although the standard error associated with predictions of permafrost less that 30 m thick is small in absolute terms (2 m), this error is significant for predictions of very thin permafrost. For predictions of permafrost more than a few metres thick, however, the Hybrid Permafrost Prediction Model satisfactorily reproduces the predictions of permafrost thickness generated by the finite-element heat conduction model.

APPLYING THE HYBRID PERMAFROST PREDICTION MODEL TO PERMAFROST MAPPING AT REGIONAL SCALES

Conventional permafrost mapping techniques rely on air-photo interpretation methods that associate various terrain characteristics (particularly vegetation cover) with the occurrence of permafrost. These methods are of limited utility for predicting permafrost thickness, or for predicting permafrost response to a changing climate. Assuming that it is possible to assign appropriate values to the various model parameters, the Hybrid Permafrost Prediction Model provides a rigorous method for predicting permafrost occurrence and thickness in a manner that is sensitive to changes in any of the model inputs. This presents an opportunity to investigate current and

possible future distributions of permafrost regionally, by using existing maps as primary data sources from which model parameters may be derived. Since Hybrid Permafrost Prediction Model equations are GIS-compatible, multiple scenarios may be modelled quickly and efficiently.

Deriving model parameter values from existing maps

Because each of the terms in the Hybrid Permafrost Prediction Model equations can be evaluated for individual sites, maps of permafrost occurrence and thickness can be produced. Existing maps of surficial geology, vegetation, and organic cover, and Digital Elevation Models (DEMs) constructed from NTS contour maps for each study area serve as primary data sources. All data were resolved to a 30 m x 30 m grid-cell format suitable for analysis within a raster-based GIS, such that each 30 m grid cell was assigned a complete set of model parameter values.

Maps of surficial geology and organic cover provided a basis for defining soil physical properties such as mineralogy, texture, and dry bulk density (e.g. Fig. 2, for Fort Simpson). Vegetation cover (e.g. Fig. 3, for Fort Simpson) was regarded as a surrogate indicator of soil water content (Jorgenson and Kreig, 1988), based on the assumption that particular vegetation communities thrive within a certain range of soil moisture conditions. Consequently, a particular species or assemblage of species will tend to dominate a given site in response to local soil moisture conditions. Parameters representing soil physical properties and water contents were used as input to Johansen's (*in* Farouki, 1981) empirical equations for calculating unfrozen and frozen thermal conductivities. Seasonal n-factor values were assigned to each vegetation unit based on information provided by Jorgenson and Kreig (1988) and Lunardini (1981).

The current climate in each area was assumed to be represented by temperature records kept at the Atmospheric Environment Service weather stations at Fort Simpson and Norman Wells. These data were converted to thawing and freezing degree-day indices based on monthly means of 30 year air temperature normals. In order to derive a value for P at each grid cell location, equivalent latitudes for individual sites were calculated on the basis of local terrain slope and aspect information derived from Digital Elevation Models of each region.

Implementing the Hybrid Permafrost Prediction Model within a GIS

Implementation of the Hybrid Permafrost Prediction Model within a geographic information system is summarized in Figure 8. Maps of predicted permafrost occurrence for present climate conditions and two warming scenarios (Fig. 9a, b, c, 11a, b, c) were produced through a simple classification of R-values generated by the Jorgenson and Kreig (1988) equation. R-values greater than 1.05 are assumed to indicate unfrozen ground. A marginal class (R-values between 0.95 and 1.05) implies that ground may be unfrozen or contain

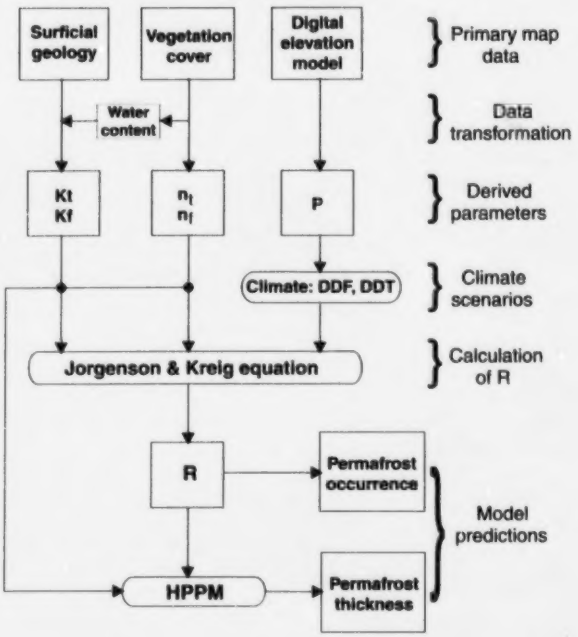


Figure 8. *Flow diagram for Hybrid Permafrost Prediction Model (HPPM) implementation. DDT = thawing degree days; DDF = freezing degree days.*

warm, thin permafrost. R-values between 0.85 and 0.95 (frozen 1) suggests permafrost occurrence, though much of this will likely disappear in response to climate warming. Permafrost classed as frozen 2 (R-values less than 0.85) will probably persist under predicted climate warming scenarios. Maps of permafrost thickness (Fig. 10a, b, c, 12a, b, c) are produced by applying the Hybrid Permafrost Prediction Model equations (Wright, 1995) to the value of R derived for each grid cell.

RESULTS AND DISCUSSION

Predictions of the aerial extent of permafrost occurrence at Fort Simpson and Norman Wells agree favourably with observations of permafrost occurrence based on interpretations of a continuous geophysical survey along the Norman Wells pipeline right-of-way conducted by Interprovincial Pipelines Ltd. (Interprovincial Pipe Lines (NW) Ltd., 1982; Kay et. al., 1983). These indicate that within the Fort Simpson study area, approximately 20% of the right-of-way is frozen. The Hybrid Permafrost Prediction Model predicts that 27% of the study area is frozen (Fig. 9a), with the marginal class accounting for another 8%. Borehole data along the pipeline corridor (Interprovincial Pipe Lines (NW) Ltd., 1982) suggest that black spruce forests in the Fort Simpson region tend to be situated on ground frozen to depths greater than 10 m, while ground underlying white spruce forest may be unfrozen, or frozen to depths less than 8 m, depending on soil characteristics. The ground beneath pine forests was usually unfrozen while under deciduous cover permafrost was almost

always absent. Additional borehole data for organic terrain units at Fort Simpson (Rutter et al., 1973) confirm that peat bogs tend to be frozen to depths of 6–8 m or more, and that fens are generally unfrozen within the top 4–6 m. Few data are available regarding thermal characteristics of fens at greater depths. These borehole data are in general agreement with Hybrid Permafrost Prediction Model predictions of permafrost thickness at Fort Simpson under current climate conditions (Fig. 10a).

At Norman Wells, mean annual air temperatures are about 2°C lower than at Fort Simpson. The Hybrid Permafrost Prediction Model predicts that approximately 81% of the region will be underlain by permafrost (Fig. 11a; exposed bedrock was excluded from the analysis). Geophysical surveys indicate that frozen terrain was encountered over 83.7% of the pipeline right-of-way (Interprovincial Pipe Lines (NW) Ltd., 1982). Studies by Judge (1973, 1975) indicate that permafrost thickness in the Norman Wells region is generally less than

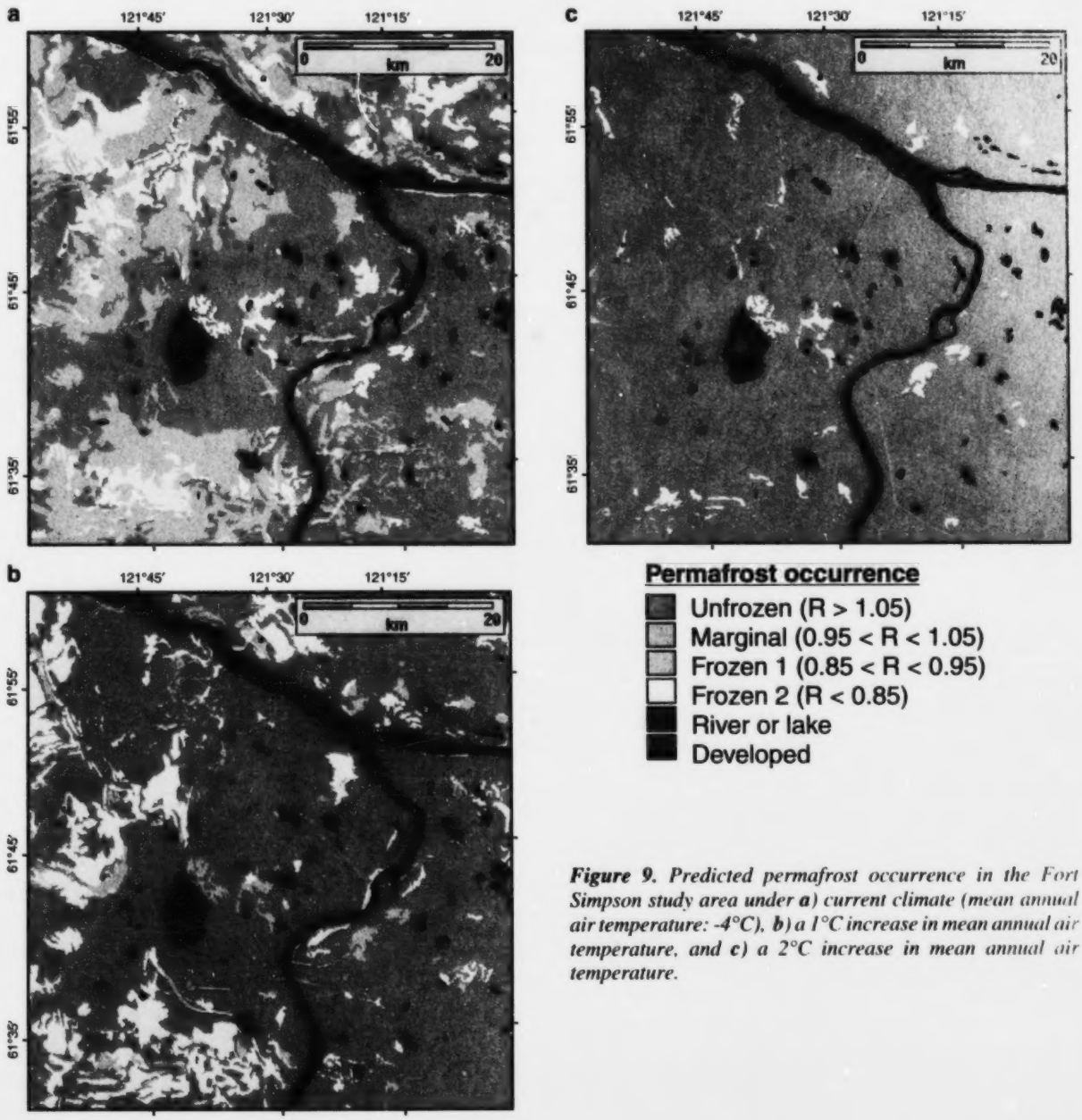


Figure 9. *Predicted permafrost occurrence in the Fort Simpson study area under* ***a****) current climate (mean annual air temperature: -4°C),* ***b****) a 1°C increase in mean annual air temperature, and* ***c****) a 2°C increase in mean annual air temperature.*

50 m and not more than 80 m. The Hybrid Permafrost Prediction Model predicts that 15% of permafrost occurring at Norman Wells is less than 20 m thick, 80% is between 20 m and 50 m, and only 5% is thicker than 50 m (Fig. 12a).

While Hybrid Permafrost Prediction Model predictions are in good agreement with geophysical and borehole data, it should be stressed that these comparisons relate to statistical evaluations of the aerial extent of permafrost along the pipeline right-of-way only, and to a small number of thickness measurements at specific borehole sites. Across most of each study area, few data are available regarding the actual occurrence and thickness of permafrost (for a more detailed evaluation of model performance, *see* Wright (1995)). Furthermore, given the relatively coarse spatial resolution of map data available at small scales, Hybrid Permafrost Prediction Model predictions provide a generalized description of permafrost characteristics within the study areas. Ground thermal conditions at any particular site must be evaluated in terms of the probability or likelihood of permafrost occurrence.

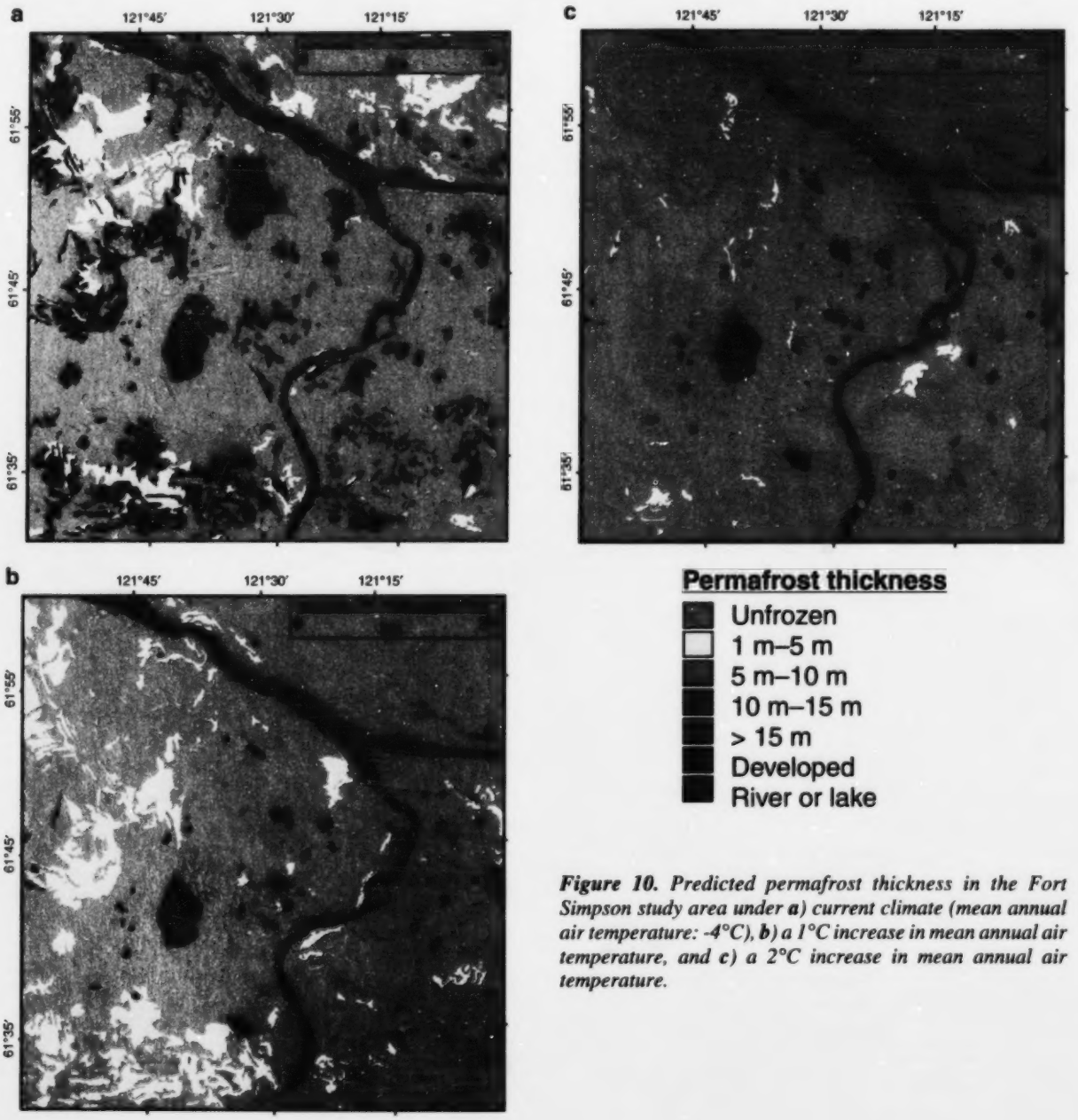


Figure 10. *Predicted permafrost thickness in the Fort Simpson study area under* ***a)*** *current climate (mean annual air temperature: -4°C),* ***b)*** *a 1°C increase in mean annual air temperature, and* ***c)*** *a 2°C increase in mean annual air temperature.*

If these predictions are accepted as reasonable representations of permafrost distribution under current climate conditions, the Hybrid Permafrost Prediction Model may be used to investigate various climate warming scenarios by rerunning the model using degree-day inputs which reflect the alternate temperature regimes (in this case, for 1°C and 2°C increases in the mean annual air temperature). Figures 9b, 9c, 10b, and 10c show Hybrid Permafrost Prediction Model predictions of permafrost occurrence and thickness at Fort Simpson under these two warming scenarios. Under a 1°C warming, the aerial extent and thickness of permafrost is substantially reduced, whereas permafrost is predicted to nearly disappear in the 2°C warming scenario. By contrast, at Norman Wells permafrost persists under both warming scenarios, but Hybrid Permafrost Prediction Model predictions suggest that its equilibrium thickness would be reduced considerably (Fig. 11b, c, 12b, c).

The Hybrid Permafrost Prediction Model exhibits a promising potential as a tool for investigating the regional characteristics of permafrost under current climate conditions. Because the model has been derived from physical relations between parameters representing the various climate and terrain elements considered to be most important in the genesis and maintenance of frozen ground, it may be applied to investigations of permafrost in different regions, or under alternative climate scenarios. This Hybrid Permafrost Prediction Model includes all the possibilities of static models

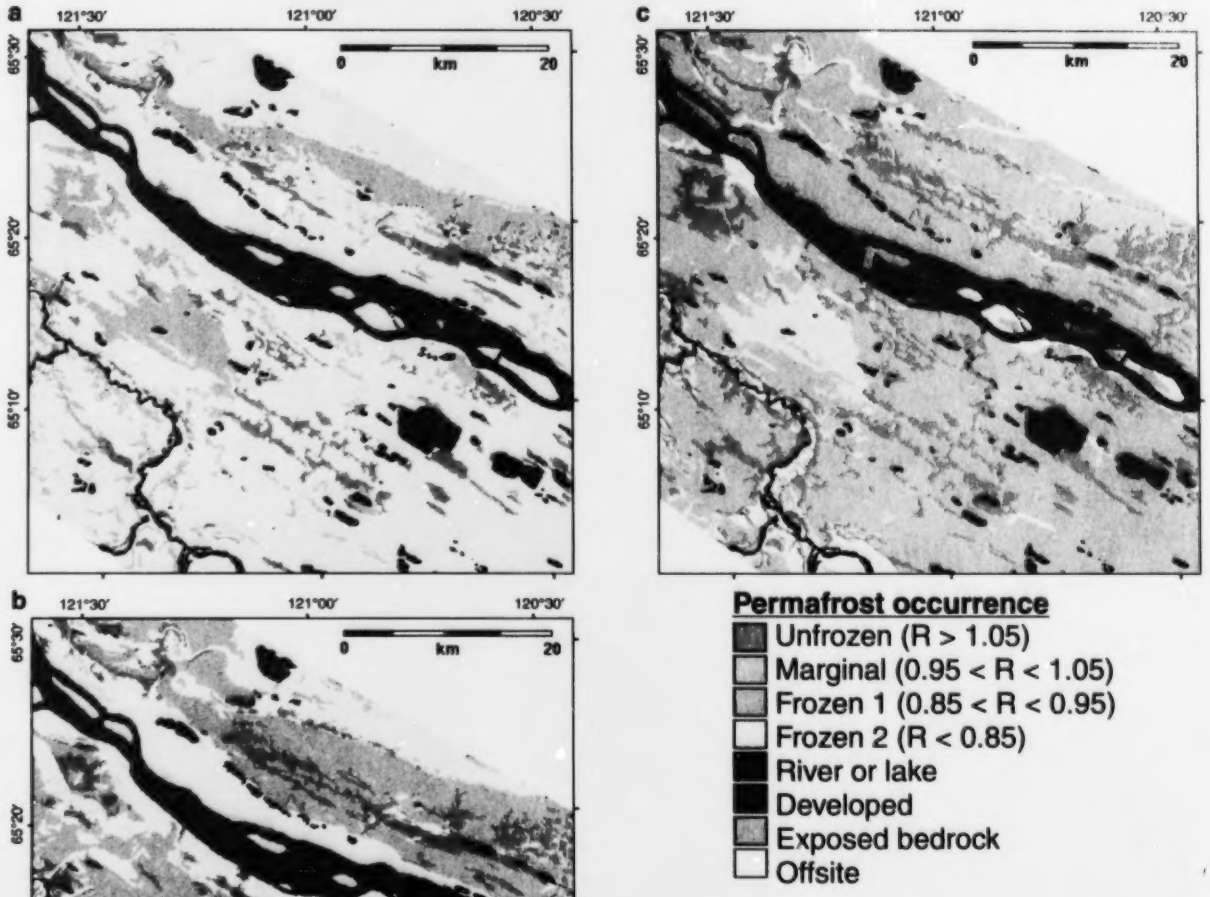


Figure 11. *Predicted permafrost occurrence in the Norman Wells study area under* ***a****) current climate (mean annual air temperature: -6°C),* ***b****) a 1°C increase in mean annual air temperature, and* ***c****) a 2°C increase in mean annual air temperature.*

that simply associate local vegetation and/or terrain characteristics with the presence or absence of permafrost. In addition, it allows for the prediction of permafrost thickness and the investigation of permafrost responses to various scenarios of climate change; however, successful application of the Hybrid Permafrost Prediction Model depends on the degree to which available climate and terrain data can be transformed into reasonable values for model parameters. Further work in this regard should be undertaken before the Hybrid Permafrost Prediction Model can be applied with confidence to regional-scale permafrost studies.

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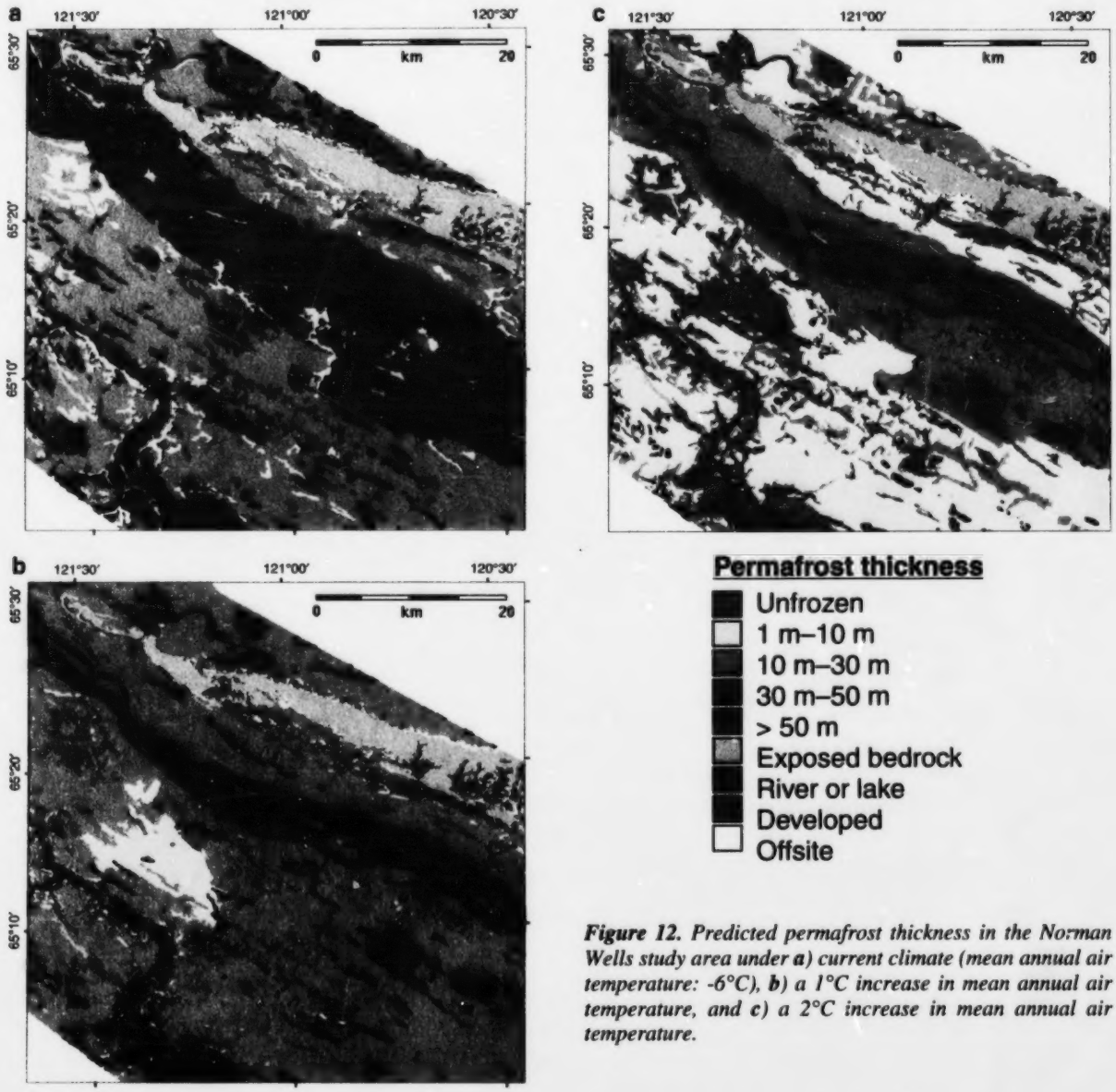

Figure 12. *Predicted permafrost thickness in the Norman Wells study area under* ***a****) current climate (mean annual air temperature: -6°C),* ***b****) a 1°C increase in mean annual air temperature, and* ***c****) a 2°C increase in mean annual air temperature.*

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